HYDROCLIMATE IN EURASIA FROM THE ARCTIC TO THE TROPICS

BY

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Abstract

Hydrometeorology in Eurasia connects the Arctic with lower latitudes through exchanges in moisture and teleconnections influencing climate variability. This thesis investigates the role of dams on the Kolyma basin, of precipitation and temperature change on a pristine permafrost lined basin of the Yana, and of changing snow cover over Eurasia on the Indian Monsoon. These three pieces of work illustrate different aspects of a changing climate that impact Eurasian hydrometeorological variations.

The Kolyma is one of the large rivers which flows into the Arctic Ocean where there has been a large winter increase and summer decrease in flow over the 1986-2000 period. Winter months are characterized by low flow while summer months by high flow. Reservoir regulation was identified as the main cause of changes in the discharge pattern, since water is released in winter for power generation and stored in summer for flood control. The overall discharge to the Arctic Ocean has decreased for Kolyma basin, despite the increase during winter. This study documents how human activities (particularly reservoirs) impact seasonal and regional hydrological variations.

The Yana Basin is a small pristine basin that has experienced minimal human impact and is ideal for investigating the role of climate variability on discharge. The precipitation discharge and temperature discharge analysis for Ubileinaya suggests that increased precipitation and higher temperatures resulted in higher discharge, but other parameters also come into play since greater precipitation does not always yield higher discharge. Overall our analysis for this station has increased our understanding of natural basins and how the climate variables like precipitation and temperature play a role.

Recent increases in May-June Indian monsoon rainfall were investigated in the context of Eurasian snow cover variations since the onset of the monsoon has long been linked to Himalayan snow cover. Himalayan snow cover and depth have decreased and this study argues that this is the driver of increased rainfall during May-June, the pre-monsoon and early monsoon period. In addition, there has been an increase in snow water equivalent in Northern part of Eurasia and decrease in Southern part, suggesting that the anomalies are large-scale. Storm track analysis reveals an increase in the number of storms in northern and a decrease in southern Eurasia. The large-scale Eurasian snow increases have been shown by other studies to be linked to Arctic sea ice decline. The direct linkage between fall Arctic sea ice decline and an increase in May-June Indian monsoon rainfall is proposed in this work but the exact climate mechanism is tenuous at this point.

This study is focused on understanding changing Arctic rivers and the connection of the Arctic with the Indian monsoon. Our study has shed some light into the connection between the Arctic and the tropics. This study could benefit from modeling study where we could have case study with and without sea ice to understand better how that could impact the monsoon and the hydrological cycle in the present and the future. Better understanding of the mechanism would help us take steps towards better adaptation policies.

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There are three papers as part of this thesis. I would like to acknowledge all the people who have helped materialize these papers

Dr. Daqing Yang conceived the presented idea about the Kolyma paper. I performed the computation and wrote the paper. Dr. Yang verified the analytical methods and supervised the findings of this work. All authors discussed the results and contributed to the final manuscript.

The Yana paper was an extension of the Kolyma paper where it was important to understand how climate variables were impacting the Arctic River. Dr. Daqing Yang developed the theory behind the paper. I performed the computation and wrote the paper. Dr. Yang verified the analytical methods and supervised the findings of this work.

Dr. Uma Bhatt devised the project ‘Link Between Changing Indian Monsoon Seasonality and the Cryosphere, the main conceptual ideas and proof outline. Dr. Uma Bhatt also supervised the project. I performed the data analysis, did the coding to plot the figures and Dr. Soumik Basu helped me with the coding and discussion throughout the formation of the paper. Dr. Peter Bieniek provided the storm track processed data for the data analysis. Dr. Shekhar provided the ground truth data for two Himalayan stations. Dr. Krishnamurthy is an expert in Monsoon, he provided the guidance throughout the development of this paper. Dr. Uma Bhatt verified the results, the computation and supervised the findings of this paper. Each author’s contribution has been critical in devising this paper.

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1. Introduction

1.1 Introduction to Arctic Hydrology

Future Arctic climate states depend on anthropogenic emissions as well as natural climate variability. The Arctic is undergoing a rapid transition, at rates that are significantly faster than the global average and its hydrological system is responding to this change through the change in river discharge, sea ice, precipitation and permafrost among many other factors (Figure 1.1) (IPCC 2014). The minimum summer sea ice extent has continued to shrink in recent decades, and the Arctic Ocean is projected to become nearly ice-free in summer by the end of this century. Not all the changes in the cryosphere are marked by declines. The duration of snow cover extent and snow depth are decreasing in North America but are increasing in Eurasia from 1880-2012 (IPCC 2014) (Figure 1.2). Rivers may be seen as integrators of changes in the hydrological cycle thus making changes more easily detectable.

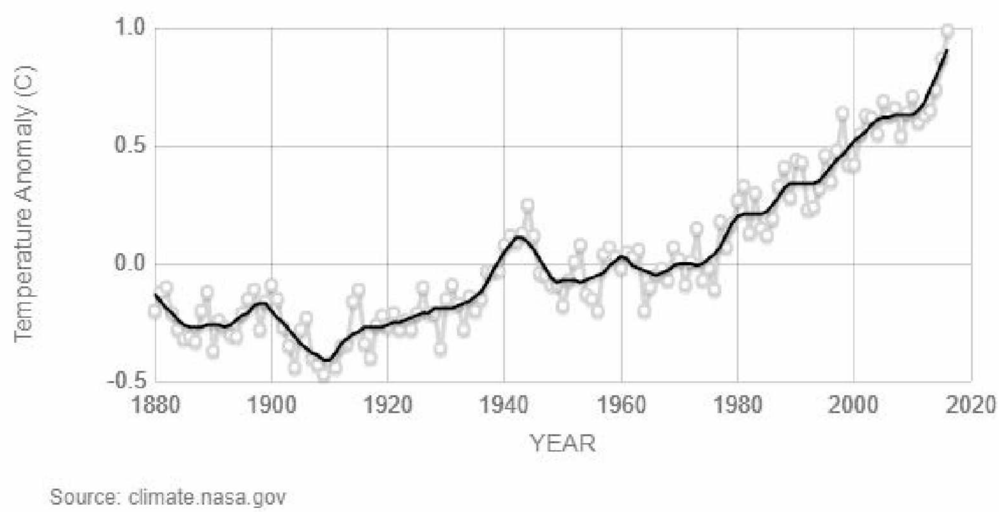

Figure 1.1. Increase in temperature from 1880 to future projection till 2020.

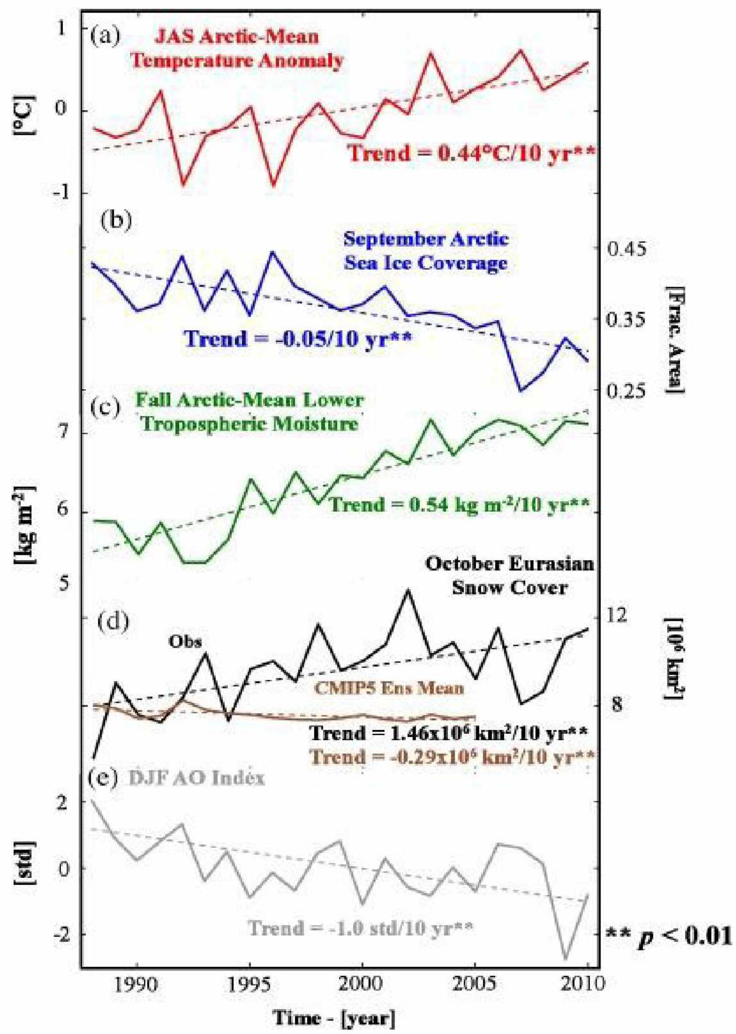

Figure 1.2. This figure has been taken from Cohen et al. 2012 (a) JAS area-averaged (poleward of 60◦N) surface temperature anomalies (◦C) from NASA MERRA. (b) September area-averaged (poleward of 65◦N) Arctic Ocean sea ice coverage (fractional area). (c) September–October vertically integrated (700–1000 hPa) and area-averaged (poleward of 60◦N) specific humidity (kg m^{-2}). (d) October mean snow cover areal extent (106 km^2) over the Eurasian continent from observations (black) and the ensemble mean from the historical runs of the CMIP5 model output (brown line). (e) The DJF average AO index (standardized). Same-colored dashed lines in (a)-(e) represent the linear trend in each index. A double asterisk (**) indicates trends that are significant at the $p < 0.01$ level.

Large rivers in the Arctic can transport vast amounts of heat across large continental watersheds in the Arctic Ocean (Nghiem et al. 2014) and the recent warming of McKenzie river temperatures have reduced sea ice in the Beaufort sea. Changes in the freshwater fluxes impact the Arctic Ocean dynamics and are part of the global ocean-atmosphere. As river discharge to the Arctic increases and this additional freshwater enters the North Atlantic, it has the potential to disrupt deep convection and thereby inhibit the global thermohaline circulation. The thermohaline circulation is an important process by which ocean currents redistribute heat and help moderate the climate (Alkire et al. 2015).

Recent studies analyzed 19 circumpolar rivers between 1977–2007 and found an increase in annual flow by 9.8% (Overeem and Syvitski 2010) accompanied by shifts in flow timing. May snowmelt has led to increased May discharge but decreased peak discharge (Overeem and Syvitski 2010). A study of 64 Canadian rivers shows a decreasing trend from 1964-2003 (Déry and Wood 2005), (Figure 1.3(e)). Many studies document an increase in winter discharge especially in Eurasia, with a decrease in North America and unchanged flow in small basins of eastern Eurasia (Rennermalm et al. 2010). There is growing evidence that the global and pan-Arctic hydrological cycles have intensified (Huntington 2006; Holland et al. 2007; Zhang et al. 2012). Different aspects of the pan- Arctic freshwater system have been impacted which include enhanced atmospheric moisture transport from lower to higher latitudes (McClelland et al. 2006), more frequent hydrological extremes (Tebaldi et al. 2006), and increasing river discharge to the Arctic Ocean (McClelland et al. 2006). The rapidly declining sea ice is expected to further alter the global hydrological cycle.

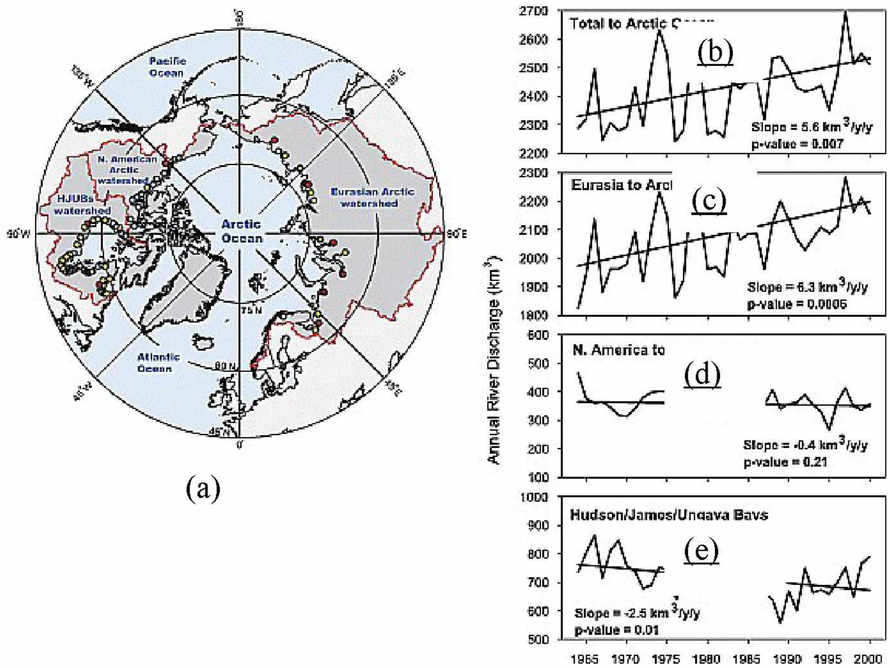

Figure 1.3. (a) Map showing watersheds of the Arctic Ocean and Hudson James, and Ungava Bays (HJUBS), with discharge trends for rivers draining these different watersheds shown in the accompanying graphs (b, c, d, e) (White et al. 2007).

In addition to climate change, anthropogenic activities such as reservoir constructions, inter-basin water diversions, and water withdrawal for industrial and agricultural uses also affect river discharge regimes and changes over space and time (Adel 2002; Vörösmarty et al. 1997). We investigate the role of anthropogenic activities and a pristine basin in our study to better understand the changes in discharge and its implication. The purpose of this study is to help begin to fill some of these gaps in knowledge of the fundamental, physical mechanisms that drive and characterize the Arctic discharge regime while drawing links to components of the earth system. Further efforts are needed to study hydrologic responses to climatic changes and human influences in the high latitudes (Shiklomanov 1978; Vörösmarty and Sahagian. 2000). Thus, there is an urgent need to better understand the role of climate variability, climate change, and anthropogenic disturbances on Siberian Rivers which form the vital links between the atmosphere, Arctic, and the Arctic Ocean. Climate change and other factors may alter these natural pathways for freshwater, leading to further environmental and societal change in the Arctic and the tropics (Déry et al. 2009). This thesis will assess three major topics in the context of anthropogenic impact, climate variability, and change in the Eurasian Arctic. The first two papers are studies that have focus on the hydrology of two Siberian rivers. The Kolyma river is analyzed to understand the role of dams in river discharge and the Yana river focuses on the change in river discharge related to temperature and snowmelt. Rivers in the Arctic play a significant role in the global climate system by contributing a large amount of discharge into the Arctic Ocean. Humans can directly affect land surface hydrology by redistributing runoff over space and time, through man-made structures such as dams. These structures can alter the natural partitioning of precipitation between runoff, evapotranspiration, and storage changes, such as via reservoir construction, irrigation, or land use changes.

1.2 Introduction to Indian Monsoon and its link to the Arctic

Precipitation is essential to sustain life on this planet and is an important parameter to assess the impact of a changing climate on agriculture production in India (Kishore et al. 2016). The Indian monsoon significantly affects agricultural planning which accounts for approximately 17% of India's GDP and approximately 49% of employment (Chatterjee et al. 2009). Precipitation over the Indian subcontinent is highly variable in time and space, leading to large-scale floods and droughts. A slight deviation in the onset of Monsoon may lead to severe droughts or floods (Figure 1.4) causing damage to infrastructure and crop loss which is devastating to a large segment of the population (Kishore et al. 2016). Therefore, the study of changes in the spatial and temporal distributions of rainfall in India has great relevance in the context of planning and policy formulation (Dash et al. 2009).

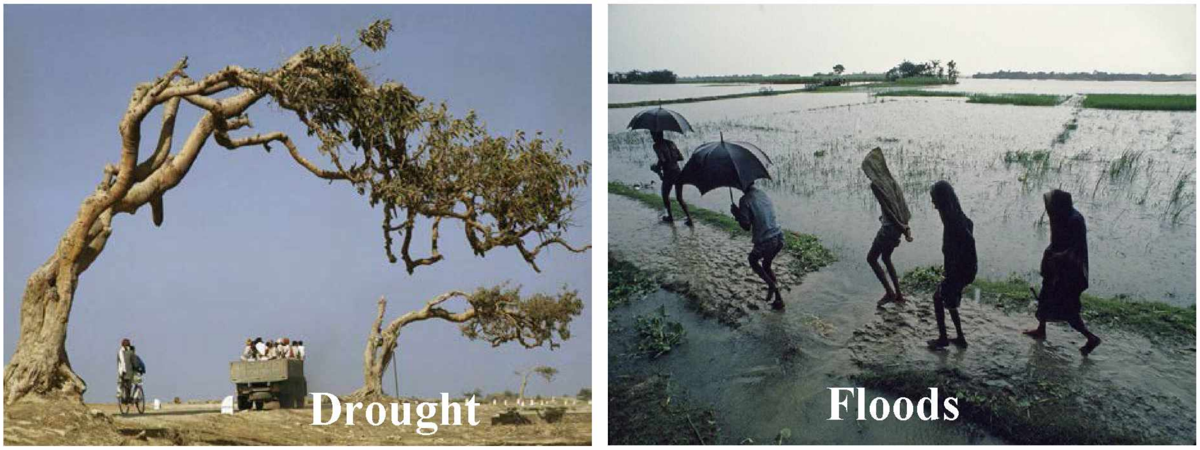

Figure 1.4. The extreme fluctuations of the Indian monsoon leads to drought or flood conditions, which impact millions of people in the Indian Subcontinent (Photos courtesy of: National Geographic).

Precipitation in the pre-monsoon month of May and monsoon month of June over India is highly variable and in recent years there has been an increase in May-June rainfall and decrease in July-August rainfall (Figure 1.5). Enhancing the ability to predict the Indian monsoon has great benefit and has been an area of research since the 1800s. Blanford (1884) was the first to demonstrate that the amount of Himalayan snow cover amount influenced the land surface's thermal characteristics which in turn altered the onset of the Asian summer monsoon. His proposed mechanism can be described as follows. When snow is extensive, the monsoon is weaker than average due to a reduced land-sea contrast and when the snow cover is below average, the monsoon is stronger than average. Bamzai and Shukla (1999) found that western Eurasia winter snow cover has a significant inverse correlation Indian summer monsoon rainfall. The mechanism was similar to Blanford with snow reducing the land-sea contrast.

This research aims to investigate whether changes in Eurasian snow cover provide an explanation for the recent increases of May-June rainfall (Fig. 1.5), which can be described as a shift in the seasonality of the monsoon. Finally, this study explores a more speculative link between sea ice and spring monsoon through Eurasian snow cover. Studies have shown that Arctic sea ice decline has been linked to an increase in snow depth in parts of Siberia (Ghatak et al. 2012). In contrast, the western part of Himalayas shows a decrease in snow depth (Shekhar et al. 2010). This study investigates whether this decrease in snow depth could lead to higher May-June rainfall. Several studies suggest that the variability of Arctic sea ice extent (Yamamoto et al. 2006; Francis et al. 2009; Honda et al. 2009; Zhang et al. 2011; Frankignoul et al. 2014; Landerer et al. 2010; Koenigk et al. 2016; King et al. 2016) and Eurasian snow cover extent (Cohen and Entekhabi 1999; Cohen et al. 2007; Cohen and Rind 1991; Cohen and Jones 2011) have some influence on the atmosphere during winter over Eurasia. The influence of Sea Ice (SIC) on Eurasian Snow Cover Extent (SCE) is thought to be strongest in fall (October and November). Studies by Cohen et al. (2014) and Wegmann et al. (2015) show that continental SCE and Arctic SIC are linked, as a reduced Arctic sea ice extent leads to a moistening of the atmospheric boundary layer, which increases the moisture flux into eastern Siberia to increase snowfall, (Luo et al. (2016) showed that reduced sea ice is connected with Ural blocking, which is associated with warm Arctic and cold Eurasian anomalies during winter. However, only a few studies have investigated links between the SCE and sea ice and while somewhat speculative, it is explored in this thesis.

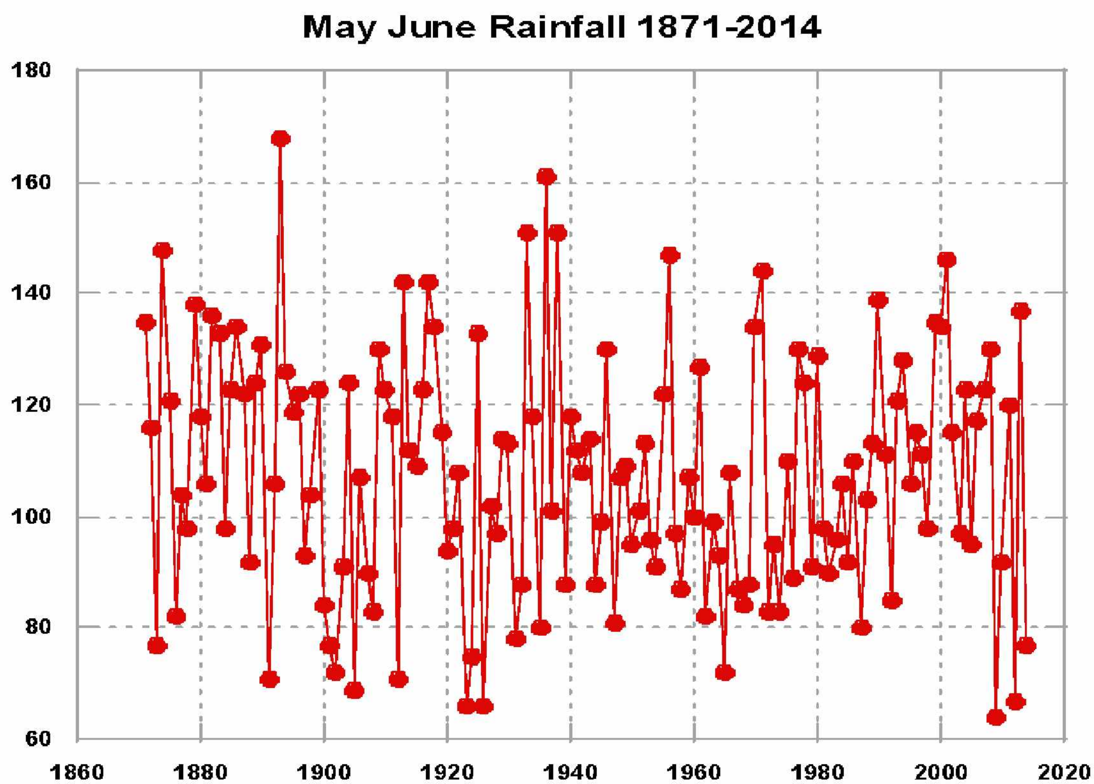

Figure 1.5. Variability of May June rainfall over 100 years of instrumental data.

The relative impact on the atmosphere of Arctic sea ice and Eurasian snow cover is largely unknown. In this study, we investigate the reason behind the increase in May-June rainfall and its possible connecting with the declining sea ice. The tropical teleconnections may both influence the snow cover over Eurasia and modify the atmospheric circulation (Fasullo 2004), leading to a possible confusion between cause and effect. Numerous studies have investigated the inverse relationship between Eurasian snow and Indian summer monsoon from June-September (Hahn and Shukla 1976; Dey and Bhanu Kumar 1982; Dickson 1984; Shukla and Mooley 1987; Khandekar 1991; Parthasarathy and Yang 1995; Bamzai and Shukla 1999; Kripalani et al. 2003). Several modeling studies have also investigated these mechanisms (Shukla and Mooley 1987; Barnett et al. 1989; Vernekar et al. 1995; Douville and Royer 1996;

Bamzai and Marx 2000; Gong 2004; Dash et al. 2005; Turner and Slingo 2011; Saha et al. 2013). Some studies based on statistical analysis (Shinoda 2001; Robock et al. 2003; Peings and Douville 2010) have questioned the validity of this teleconnection and argue that the El Niño–Southern Oscillation (ENSO) modulates the Indian monsoon. Other mechanisms have also been suggested, Meehl and Washington (1993) suggested that surface temperature anomalies over Eurasia and South Asia are controlled by changes in large-scale mid-latitude circulation that are, in turn, driven by convective heating anomalies associated with the tropical biennial oscillation (TBO). Fasullo (2004) found that the inverse relationship between Eurasian snow cover and rainfall mainly over northern parts of India to be stronger during neutral ENSO years. In contrast, modeling studies (Ferranti and Molteni 1999; Turner et al. 2007) found that this relationship still holds and is independent of ENSO-ISMR relationship. Senan et al. (2016) have shown that years with higher snow depth over the Himalayan-Tibetan Plateau region in early April can delay the onset of the monsoon. However, none of the above studies have investigated in detail the impact of the cryosphere on May June rainfall and its variability.

The link between the Arctic and the Indian monsoon across Eurasia has received some attention. Buermann et al. (2005) used statistical techniques and concluded that springtime surface conditions in India are a probable "bridge" between the wintertime Arctic circulation and summer monsoon rainfall during June in the northwestern part of India. Studies have looked at the influence of spring Arctic sea ice on the East Asian summer monsoon (EASM; Guo et al. 2014). Many observational and model-based studies suggest that Arctic sea ice changes impact regional and remote climate (Budikova 2009). Zhao et al. (2004) connected spring sea ice cover in the Bearing Sea and the Sea of Okhotsk and rainfall variability in the East Asian summer monsoon (EASM). Wu, Zhang, and Wang (2009) found a significant correlation between the spring Arctic sea ice concentration (ASIC) and summer rainfall over China using observational and reanalysis data for the last 30 years.

It is important to understand the implications of rapidly decreasing sea ice and its association with changes in surface hydrology over the Eurasian region and whether it could also affect the seasonality of the Indian monsoon. The main purpose of the present study is to investigate a new possible mechanism where fall Arctic sea ice impacts the winter and spring snow in Eurasia and Himalaya, which changes the surface conditions in the northern part of India leading to enhanced rainfall in May-June month.

1.3 Scientific Questions and Objectives

The overall goal of this thesis is to understand the hydrological changes in Eurasian with an attempt to link Arctic the tropics. Much is generally known about the impacts of teleconnections on Arctic hydrology and much still needs to be understood about possible links between the Eurasian Arctic to the tropics. However, the roles of anthropogenic impacts need to be understood along with how the climate variability plays into the changing seasonality of monsoon. Elucidating these mechanisms is necessary to better understand climate variability and change in the Arctic and its global implications. This thesis will investigate three topics in the area of Eurasian hydrology based on the analysis of observed data and global climate model simulations.

Paper 1: Streamflow characteristics and changes in Kolyma Basin in Siberia

- Is there a change in seasonality of the flow?
- Is the winter base flow increasing and if so why?
- Has the total annual discharge into the ocean increased or decreased?

- What impact does a dam have on the Arctic discharge?

Hypothesis: We hypothesize that since dam tends to store water in summer that the peak discharge will decrease, but the winter discharge will increase due to winter usage of the dam.

Paper 2: Streamflow analysis for the Yana basin in eastern Siberia

- How is the hydrology of a pristine basin different than the one impacted by anthropogenic activities?
- How does temperature variability impact discharge flow?
- How does precipitation variability impact the discharge?

Hypothesis: We hypothesize that warmer summers will result in greater discharge, and years with more snow will have higher peak discharge.

Paper 3: Does the Eurasian cryosphere play a role in recent increases in May-June Indian monsoon rainfall?

What are the climate drivers of the increase in May-June rainfall?

- Could variations in Himalayan and Eurasian snow be responsible for the increase in May- June rainfall?
- Is Himalayan snow variations part of larger-scale variations of snow in the Arctic? of Arctic sea ice?

Hypothesis: We hypothesize that increase in May-June rainfall is directly linked decreased snow depth in the Himalaya and indirectly linked to increased snow cover in northern Eurasia. In this thesis, we investigate changes in the Arctic and possible far-field impacts of these changes by investigating several components of the hydrological system: river discharge, snow cover, and monsoon rainfall. An attempt is made to link tropical and Arctic processes to show that the global climate is interconnected.

1.4 References

Adel, M. M., 2002: Man-made climatic changes in the Ganges basin. *Int. J Climatol.*, **22**, 993–1016, doi:10.1002/joc.732.

Alkire, M. B., A. D. Jacobson, G. O. Lehn, R. M., 2015: Small rivers could have big impact on Arctic Ocean. *Eos (Washington. DC)*.

Bamzai, A. S., and J. Shukla, 1999: Relation between Eurasian snow cover, snow depth, and the Indian summer monsoon: An observational study. *J. Clim.*, **12**, 3117–3132, doi:10.1175/1520-0442(1999)012<3117:RBESCS>2.0.CO;2.

Bamzai, A. S., and L. Marx, 2000: COLA AGCM simulation of the effect of anomalous spring snow over Eurasia on the Indian summer monsoon. *Q. J. R. Meteorol. Soc.*, **126**, 2575–2584, doi:10.1002/qj.49712656811. http://doi.wiley.com/10.1002/qj.49712656811.

Barnett, T. P., L. Dümenil, U. Schlese, E. Roeckner, and M. Latif, 1989: The Effect of Eurasian Snow Cover on Regional and Global Climate Variations. *J. Atmos. Sci.*, **46**, 661–686.

Blanford, H. F., 1884: On the Connexion of the Himalaya Snowfall with Dry Winds and Seasons of Drought in India. *Proc. R. Soc. London*, **37**, 3–22, doi:10.1098/rspl.1884.0003.

Budikova, D., 2009: Role of Arctic sea ice in global atmospheric circulation: A review. *Glob. Planet. Change*, **68**, 149–163, doi:10.1016/j.gloplacha.2009.04.001.

Buermann, W., B. Lintner, and C. Bonfils, 2005: A wintertime Arctic Oscillation signature on early-season Indian Ocean monsoon intensity. *J. Clim.*, **18**, 2247–2269, doi:10.1175/JCLI3377.1.

Chatterjee, S., S. Ghosh, and U. K. De, 2009: Reduction of number of parameters and forecasting convective developments at Kolkata (22.53°N, 88.33°E), India during pre-monsoon season: An application of multivariate techniques. *Indian J. Radio Sp. Phys.*, **38**, 275–282.

Cohen, J., and D. Rind, 1991: The Effect of Snow Cover on the Climate. *J. Clim.*, **4**, 689–706, doi:10.1175/1520-0442(1991)004<0689:TEOSCO>2.0.CO;2. http://journals.ametsoc.org/doi/abs/10.1175/1520-0442(1991)004%3C0689:TEOSCO%3E2.0.CO;2.

Cohen, J., and D. Entekhabi, 1999: Corrections to "Eurasian snow cover variability and Northern Hemisphere climate predictability." *Geophys. Res. Lett.*, **26**, 1051, doi:10.1029/1999GL900200.

Cohen, J., and J. Jones, 2011: A new index for more accurate winter predictions. *Geophys. Res. Lett.*, **38**, doi:10.1029/2011GL049626.

Cohen, J., M. Barlow, P. J. Kushner, and K. Saito, 2007: Stratosphere-troposphere coupling and links with Eurasian land surface variability. *J. Clim.*, **20**, 5335–5343, doi:10.1175/2007JCLI1725.1.

Cohen, J. L., J. C. Furtado, M. Barlow, and V. A. Alexeev, 2012: Asymmetric Seasonal Temperature Trends. Geophysical Research Letters, 39: L04705, doi: 10.1029/2011GL050582.

Cohen, J. and coauthors,2014: Recent Arctic amplification and extreme mid-latitude weather. *Nature Geosci.*,7, 627-637.

Dash, S. K., G. P. Singh, M. S. Shekhar, and A. D. Vernekar, 2005: Response of the Indian summer monsoon circulation and rainfall to seasonal snow depth anomaly over Eurasia. *Clim. Dyn.*, **24**, 1–10, doi:10.1007/s00382-004-0448-3.

Dash, S. K., M. A. Kulkarni, U. C. Mohanty, and K. Prasad, 2009: Changes in the characteristics of rain events in India. *J. Geophys. Res. Atmos.*, **114**, doi:10.1029/2008JD010572.

Déry, S. J., and E. F. Wood, 2005: Decreasing river discharge in northern Canada. *Geophys. Res. Lett.*, **32**, 1–4, doi:10.1029/2005GL022845.

Déry, S.J., M.A. Hernandez-Henriquez, J.E. Burford, and E.F. Wood. 2009. Observational evidence of an intensifying hydrological cycle in northern Canada. *Geophysical Research Letters* 36: L13402. doi: 13410.11029/12009GL038852.

Dey, B., and O. S. R. U. Bhanu Kumar, 1982: An Apparent Relationship between Eurasian Spring Snow Cover and the Advance Period of the Indian Summer Monsoon. *J. Appl. Meteorol.*, **21**, 1929–1932, doi:10.1175/1520-0450(1982)021<1929:AARBES> 2.0.CO; 2.

Dickson, R. R., 1984: Eurasian Snow Cover versus Indian Monsoon Rainfall -- An extension of the Hahn-Shulka Results. *Q. J. R. Meteorol. Soc.*, **23**, 171–173.

Douville, H., and J. F. Royer, 1996: Sensitivity of the Asian summer monsoon to an anomalous Eurasian snow cover within the Meteo-France GCM. *Clim. Dyn.*, **12**, 449–466.

Fasullo, J., 2004: A stratified diagnosis of the Indian monsoon - Eurasian snow cover relationship. *J. Clim.*, **17**, 1110–1122, doi:10.1175/1520- 0442(2004)017 <1110:ASDOTI> 2.0.CO;2.

Ferranti, L., and F. Molteni, 1999: Ensemble simulations of eurasian snow-depth anomalies and their influence on the summer Asian monsoon. *Journal of the Royal Meteoritical Society*, **125:559**, 2597-2610.

Francis, J. A., W. Chan, D. J. Leathers, J. R. Miller, and D. E. Veron, 2009: Winter Northern Hemisphere weather patterns remember summer Arctic sea-ice extent. *Geophys. Res. Lett.*, **36**, doi:10.1029/2009GL037274.

Frankignoul, C., N. Sennéchael, and P. Cauchy, 2014: Observed atmospheric response to cold season sea ice variability in the arctic. *J. Clim.*, **27**, 1243–1254, doi:10.1175/JCLI-D-13-00189.1.

Ghatak, D., C. Deser, A. Frei, G. Gong, A. Phillips, D. A. Robinson, and J. Stroeve, 2012: Simulated Siberian snow cover response to observed Arctic sea ice loss, 1979-2008. *J Geophys. Res. Atmos.*, **117**, doi:10.1029/2012JD018047.

Gong, G., 2004: Sensitivity of atmospheric response to modeled snow anomaly characteristics. *J. Geophys. Res.*, **109**, D06107, doi:10.1029/2003JD004160. http://www.agu.org/pubs/crossref/2004/2003JD004160.shtml.

Guo, D., Y. Gao, I. Bethke, D. Gong, O. M. Johannessen, and H. Wang, 2014: Mechanism on how the spring Arctic sea ice impacts the East Asian summer monsoon. *Theor. Appl. Climatol.*, **115**, 107–119, doi:10.1007/s00704-013-0872-6.

Hahn, D. G., and J. Shukla, 1976: An apparent relationship between Eurasian snowcover and Indian monsoon rainfall. *J. Atmos. Sci.*, **33**, 2461–2462.

Holland, M. M., J. Finnis, A. P. Barrett, and M. C. Serreze, 2007: Projected changes in Arctic Ocean freshwater budgets. *J. Geophys. Res. Biogeosciences*, **112**, doi:10.1029/2006JG000354.

Honda, M., J. Inoue, and S. Yamane, 2009: Influence of low Arctic sea-ice minima on anomalously cold Eurasian winters. *Geophys. Res. Lett.*, **36**, doi:10.1029/2008GL037079.

Huntington, T. G., 2006: Evidence for intensification of the global water cycle: Review and synthesis. *J. Hydrol.*, **319**, 83–95, doi:10.1016/j.jhydrol.2005.07.003. IPCC, 2014: *IPCC 2014*. 2-26 pp.

Khandekar, M. L., 1991: Eurasian snow cover, Indian monsoon and El Niño/Southern Oscillation – a synthesis. *Atmos. - Ocean*, **29**, 636–647, doi:10.1080/07055900.1991.9649422.

King, M. P., M. Hell, and N. Keenlyside, 2016: Investigation of the atmospheric mechanisms related to the autumn sea ice and winter circulation link in the Northern Hemisphere. *Clim. Dyn.*, **46**, 1185–1195, doi:10.1007/s00382-015-2639-5.

Kishore, P., S. Jyothi, G. Basha, S. V. B. Rao, M. Rajeevan, I. Velicogna, and T. C. Sutterley, 2016: Precipitation climatology over India: validation with observations and reanalysis datasets and spatial trends. *Clim. Dyn.*, **46**, 541–556, doi:10.1007/s00382-015-2597-y.

Koenigk, T., M. Caian, G. Nikulin, and S. Schimanke, 2016: Regional Arctic sea ice variations as predictor for winter climate conditions. *Clim. Dyn.*, **46**, 317–337, doi:10.1007/s00382-015-2586-1.

Kripalani, R. H., A. Kulkarni, S. S. Sabade, and M. L. Khandekar, 2003: Indian Monsoon variability in a global warming scenario. *Nat. Hazards*, **29**, 189–206, doi:10.1023/A:1023695326825.

Landerer, F. W., J. O. Dickey, and A. Güntner, 2010: Terrestrial water budget of the Eurasian pan-Arctic from GRACE satellite measurements during 2003-2009. *J. Geophys. Res. Atmos.*, **115**, doi:10.1029/2010JD014584.

Luo, D., Y. Xiao, Y. Yao, A. Dai, I. Simmonds, C. L. E. Franzke, 2016: Impact of Ural Blocking on Winter Warm Arctic–Cold Eurasian Anomalies. Part I: Blocking-Induced Amplification. *J. of Climate*., doi:10.1175/JCLI-D-15-0611.1

McClelland, J. W., S. J. Déry, B. J. Peterson, R. M. Holmes, and E. F. Wood, 2006: A pan-arctic evaluation of changes in river discharge during the latter half of the 20th century. *Geophys. Res. Lett.*, **33**, doi:10.1029/2006GL025753.

Meehl, G. A., and W. M. Washington, 1993: South asian summer monsoon variability in a model with doubled atmospheric carbon dioxide concentration. *Science (80).*, **260**, 1101–1104, doi:10.1126/science.260.5111.1101. http://www.ncbi.nlm.nih.gov/pubmed/17806338.

Nghiem, S. V., D. K. Hall, I. G. Rigor, P. Li, and G. Neumann, 2014: Effects of Mackenzie River discharge and bathymetry on sea ice in the Beaufort Sea. *Geophys. Res. Lett.*, **41**, 873–879, doi:10.1002/2013GL058956.

Overeem, I., and J. P. M. Syvitski, 2010: Shifting discharge peaks in arctic rivers, 1977-2007. *Geogr. Ann. Ser. A Phys. Geogr.*, **92**, 285–296, doi:10.1111/j.1468-0459.2010.00395.x.

Parthasarathy, B., K. R. Kumar, and D. K. Kothawale, 1992: Indian Summer monsoon rainfall indices:1871-1990. *Meteor.Mag.*, **121**, 174–186.

Parthasarathy, B., and S. Yang, 1995: Relationships between regional indian summer monsoon rainfall and eurasian snow cover. *Adv. Atmos. Sci.*, **12**, 143–150, doi:10.1007/BF02656828.

Peings, Y., and H. Douville, 2010: Influence of the Eurasian snow cover on the Indian summer monsoon variability in observed climatologies and CMIP3 simulations. *Clim. Dyn.*, **34**, 643–660, doi:10.1007/s00382-009-0565-0.

Rennermalm, A. K., E. F. Wood, and T. J. Troy, 2010: Observed changes in pan-arctic cold-season minimum monthly river discharge. *Clim. Dyn.*, **35**, 923–939, doi:10.1007/s00382- 009-0730-5.

Robock, A., M. Mu, K. Vinnikov, and D. Robinson, 2003: Land Surface Conditions over Eurasia and Indian summer monsoon rainfall. *J. Geophys. Res.*, **108**, 17, doi:10.1029/2002JD002286.

Saha, S. K., S. Pokhrel, and H. S. Chaudhari, 2013: Influence of Eurasian snow on Indian summer monsoon in NCEP CFSv2 freerun. *Clim. Dyn.*, **41**, 1801–1815, doi:10.1007/s00382-012-1617-4.

Senan, R., and Coauthors, 2016: Impact of springtime Himalayan–Tibetan Plateau snowpack on the onset of the Indian summer monsoon in coupled seasonal forecasts. *Clim. Dyn.*, **47**, 2709–2725

Shekhar, M. S., H. Chand, S. Kumar, K. Srinivasan, and A. Ganju, 2010: Climate-change studies in the western Himalaya. *Ann. Glaciol.*, **51**, 105–112, doi:10.3189/172756410791386508.

Shiklomanov, I. A., 1978: Dynamics of anthropogenic changes in annual river runoff in the USSR. *Sov. Hydrol. Selec. Pap.*, **17**, 11–22.

Shinoda, M., 2001: Climate memory of snow mass as soil moisture over central Eurasia. *J. Geophys. Res.*, **106**, 33393–33403, doi:10.1029/2001JD000525.

Shukla, J., and D. A. Mooley, 1987: *Empirical prediction of the summer monsoon rainfall over India*.

Tebaldi, C., K. Hayhoe, J. M. Arblaster, and G. A. Meehl, 2006: Going to the extremes: An intercomparison of model-simulated historical and future changes in extreme events. *Clim. Change*, **79**, 185–211, doi:10.1007/s10584-006-9051-4.

Turner, A. G., and J. M. Slingo, 2011: Using idealized snow forcing to test teleconnections with the Indian summer monsoon in the Hadley Centre GCM. *Clim. Dyn.*, **36**, 1717–1735, doi:10.1007/s00382-010-0805-3.

Turner, A. G., P. M. Inness, and J. M. Slingo, 2007: The effect of doubled CO2 and model basic state biases on the monsoon-ENSO system. II: Changing ENSO regimes. *Q. J. R. Meteorol. Soc.*, **133**, 1159–1173, doi:10.1002/qj.83.

Vernekar, A. D., J. Zhou, and J. Shukla, 1995: The effect of Eurasian snow cover on the Indian monsoon. *J. Clim.*, **8**, 248–266.

Vörösmarty, C. J., and D. Sahagian, 2000: Anthropogenic Disturbance of the Terrestrial Water Cycle. *Bioscience*, **50**, 753, doi:10.1641/0006-3568(2000)050[0753:ADOTTW]2.0.CO;2.

Vörösmarty, C. J., K. P. Sharma, B. M. Fekete, A. H. Copeland, J. Holden, J. Marble, and J. A. Lough, 1997: The storage and aging of continental runoff in large reservoir systems of the world. *Ambio*, **26**, 210–219.

Wegmann, M., and Coauthors, 2015: Arctic moisture source for Eurasian snow cover variations in autumn. *Environ. Res. Lett.*, **10**, 54015, doi:10.1088/1748-9326/10/5/054015. http://stacks.iop.org/1748-9326/10/i=5/a=054015?key=crossref.fe92c673fbffc6cc390fb7db731e4a5a.

White, D., and Coauthors, 2007: The arctic freshwater system: Changes and impacts. *J. Geophys. Res. Biogeosciences*, **112**, doi:10.1029/2006JGO00353.

Wu, B., R. Zhang, and B. Wang, 2009: On the association between spring Arctic sea ice concentration and Chinese summer rainfall: A further study. *Adv. Atmos. Sci.*, **26**, 666–678, doi:10.1007/s00376-009-9009-3.

Yamamoto, K., Y. Tachibana, M. Honda, and J. Ukita, 2006: Intra-seasonal relationship between the Northern Hemisphere sea ice variability and the North Atlantic Oscillation. *Geophys. Res. Lett.*, **33**, doi:10.1029/2006GL026286.

Zhang, Q., C. Y. Xu, X. Chen, and Z. Zhang, 2011: Statistical behaviours of precipitation regimes China and their links with atmospheric circulation 1960-2005. *Int. J. Climatol.*, **31**, 1665–1678, doi:10.1002/joc.2193.

Zhang, X., J. He, J. Zhang, I. Polyakov, R. Gerdes, J. Inoue, and P. Wu, 2012: Enhanced poleward moisture transport and amplified northern high-latitude wetting trend. *Nat. Clim. Chang.*, **3**, 47–51, doi:10.1038/nclimate1631.

Zhao, P., X. Zhang, X. Zhou, M. Ikeda, and Y. Yin, 2004: The sea ice extent anomaly in the north Pacific and its impact on the east Asian summer monsoon rainfall. *J. Clim.*, **17**, 3434–3447.

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2. Streamflow Characteristics and Changes in Kolyma Basin in Siberia[1]

2.1 Abstract

This study documents major changes in streamflow hydrology over the Kolyma watershed due to climatic variations and human impacts. Streamflow seasonal cycles over the basin are characteristic of the northern region, with the lowest runoff in April and peak flow in June. Analyses of monthly flows and trends show that reservoir construction and operation have considerably affected streamflow regimes. Comparisons of mean monthly discharge records between pre- and post-1986 dam periods indicate that the mid-lower basin (downstream of the dam) experienced significant increase in low flows and decrease in peak flows after dam construction. For example, mean monthly flows during the post-dam period at the Ust'-Srednekan station (located 1,423 km downstream of the dam) has strongly increased by about 205 m^3s^{-1} (or 522%-3,157%) during December-April, and decreased by 133 m^3s^{-1} (41%) in June. Long-term monthly discharge data reveal an overall increase in streamflow during low flow seasons; the increase is greater for the stations located downstream of the dam. The Srednekolunsk station (1,720 km from dam) shows low flow increase ranging from 130 (43%) to 268 m^3s^{-1}(454%) during November–April, and high discharge decrease by 2,550 to 519 m^3s^{-1} during June-August in the post-dam era (1986-2000). These changes in flow patterns are mainly caused by reservoir regulation, as reservoirs release water in winter for power generation and store water in summer for flood control. Dam impact on flow regimes and changes are visible along the main river trunk; thus, the cold season discharge increase at the basin outlet is primarily the result of reservoir regulation. Annual discharge records show different changes within

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the Kolyma basin, with moderate increases in the upper basin and weak decreases in the mid–lower basin. Overall annual discharge near the basin outlet has decreased by 1.5% during 1978–2000. This study emphasizes the importance of human activities (particularly reservoirs) on seasonal and regional hydrology changes and points to the need to further examine natural causes and human impacts over other high-latitude watersheds.

2.2 Introduction

Rivers in the Arctic play a significant role in the global climate system by contributing a large amount of discharge into the Arctic Ocean. The Arctic Ocean receives 3,685 km^3 of freshwater discharge in a year, which is about 11% of annual freshwater discharge into the global oceans (Shiklomanov 2000; Shiklomanov and Shiklomanov 2003). The amount of freshwater inflow affects ocean salinity, thermohaline circulation, and sea ice formation (Aagaard and Carmack 1989). The Arctic Ocean is the most landlocked and most river-influenced of all oceans (Vörösmarty et al. 2000). River influence to the Arctic Ocean is pronounced on the shallow shelf regions, particularly in Russia (Lammers et al. 2001). The Arctic hydrological system varies over space and time because of large-scale variations in atmospheric circulation (Kane 1997; Walsh 2000; Serreze 2003). Variations and changes in the hydrological cycle at local and regional scales affect vegetation patterns (Foley et al. 1994), permafrost dynamics (Nelson and Anisimov 1993; Kane et al. 1990), and gas fluxes (Oechel et al. 1993).

Significant changes have been observed in the large Arctic river basins. For instance, Ye et al. (2003) and Yang et al. (2003, 2004 a, b) found discharge increase during the winter months and shift in peak discharge timing related to winter and spring warming over the large watersheds in Siberia. Recent studies have focused on the mechanisms driving these changes (McClelland et al. 2004). Peterson et al. (2002) suggested that the transport of moisture from the lower to higher latitudes might be responsible for river runoff increases. Berezovskaya et al. (2004) recently reported inconsistency in yearly precipitation and runoff trends for the large Siberian Rivers. A better understanding of the hydroclimate system is critical to understand and explain the observed hydrologic changes. In addition to climate change, anthropogenic activities such as reservoir constructions, inter-basin water diversions, and water withdrawal for industrial and agricultural uses also affect river discharge regimes and changes over space and time (Miah 2002; Vörösmarty et al. 1997; Revenga et al. 1998; Dynesius and Nilsson 1994). Slow economic growth and low population in the high-latitude regions have resulted in low impact by humans (Shiklomanov et al. 2000; Lammers et al. 2001). Among the human impacts, reservoir regulation has the most direct effects on hydrologic regimes and changes (Ye et al. 2003; Yang et al. 2004 a, b). More efforts are needed to study hydrologic responses to climatic changes and human influences in the high latitudes. This study systematically analyzes long-term monthly and yearly discharge for the Kolyma River watershed, with the emphasis on defining streamflow regimes and long-term changes induced by climate variations and human impacts. The results of this study will improve our understanding of hydrologic response to climate change in the Arctic regions.

2.3 Basin description, datasets, and methods of analyses

The Kolyma watershed situated in eastern Siberia is the sixth largest river flowing into the Arctic Ocean (Tsuyuzaki et al. 1999; IUCN 2006). The river is 2,410 km long; it rises in the Kolyma and the Chreskogo ranges and flows into the Arctic Ocean. The Kolyma River discharges 100.8 km^3 yr^{-1} to the Arctic Ocean. The Kolyma River is mostly fed by spring snowmelt and summer rainfall with the annual precipitation inputs to the watershed of 47% snow and 53% rain (Welp et al. 2005). The dominant land cover types in the Kolyma region are shrub (41%)

and forest (31%; Revenga et al. 1998). The basin is characterized by taiga flora (Swanson 2003). Tree lines are lower than 1,000 m, above which exists alpine tundra. The primary tree types are *Larix Mill* (larch) and *Picea A. Dietr* (spruce), and upland shrubs consist mainly of *Pinus pumila* (pine; Swanson 2003). The Kolyma basin has the typical Arctic climate of a long winter and a short summer. Temperatures are low during September to May; mean January temperatures are about -30° to - 40°C (Swanson 2003). Low temperatures promote continuous permafrost in the region (Kuchment et al. 2000). The active layer varies from 0.2 m in the tundra to 1.0 m in the taiga. The soil starts to thaw after snowmelt in late May and freezes in early September. Permafrost acts as an impermeable surface, resulting in limited exchange between surface water and subsurface permafrost (Panfilova 1986). The Kolyma region is not very developed except for the dam constructions during the late 1980s and widespread gold mining activities. Hydrological observations in the Siberian regions, such as discharge, stream water temperature, river-ice thickness, dates of river freeze-up and breakup, have been carried out since the mid-1930s by the Russian Hydro-meteorological Services, and the observational records were quality controlled and archived by the same agency (Shiklomanov et al. 2000). The discharge data are now available from the R-ArcticNet (version 3.0) - a database of pan-Arctic River discharge (online at http://www.r-arcticnet.sr.unh.edu/main.html). Within the Kolyma basin, a hydrologic observing network consisting of 13 stations has been set up since the mid-1930s. Ten stations with long- term monthly and annual discharge records (the Kolymskoye, Srednekolunsk, Yasachnaya, Ust'-Srednekan, Sinegor's, Duscania, Orotuk, Kulu, Bol'shoy Anuy, and Oloy) (Fig. 2.1) have been used for this analysis. Relevant station information is summarized in Table 1. In addition, basin-mean monthly and yearly temperature and precipitation records derived from the global (observational climatic) datasets (Jones 1994; Hulme 1991) are also utilized in this study.

The warming of permafrost is important for Arctic hydrology. As permafrost degrades surface water will become closely connected to the subsurface groundwater. As a result, soil drainage will improve and basin storage will also change. These changes will alter the streamflow seasonal cycle (Hinzman et al. 2005). Holmes et al. (2003) examined the potential role of permafrost thaw as a significant contributor to the observed discharge changes over large rivers in Siberia. They found that thawing of permafrost may contribute to changes in yearly discharge but cannot be considered as a major driver. This paper did not investigate permafrost effect since we did not have permafrost data over the Kolyma basin. To examine the hydrological regimes and changes over the Kolyma basin, we first compiled the basic geophysical and hydrological information and identified dam-regulated and unregulated sub-basins. We divided the basin into upper, middle, and lower sub-basins to better examine the regional hydrological features and the effect of dam regulation. Second, we calculated and compared long-term monthly mean discharge, standard deviation, and trend for the major discharge stations. Trends were determined by a linear regression. The total trend was defined as the difference between the first and last point on the regression line. The trends for different stations have different time periods. We limited the trend calculations and discussions for flow records longer than 15 yr. The trend was expressed in terms of total discharge change over the observation period for each station. We understand that trend results are less compatible among the stations because of varying data periods used here. Flow trends in regulated basins are not very useful to examine climate impact on regional hydrology changes (Ye et al. 2003; Yang et al. 2004 a, b). However, trend analyses and results are necessary to document flow regime changes due to regulation over time. Significance of the trend was evaluated by the standard Student's t-test. Third, we compared the mean streamflow between the pre- and post-dam periods, so as to quantify the effects of reservoir operation within the basin and at the basin outlet. To better determine the impact of dam regulation over the basin, we also analyzed discharge budget (changes) along the main river valley, and examined and compared discharge changes between the regulated and unregulated sub-basins. Finally, we discuss the relationship between monthly climate and hydrology changes over the basin as a whole.

2.4 Streamflow regime and change

This section discusses discharge regimes and changes at the 10 stations located in the main river valley and in the tributaries (Fig. 2.1). It also documents the dam in the basin and quantifies its influence on the seasonal characteristics of streamflow hydrology.

a. Kulu (upper basin)

The Kulu station (A in Fig. 2.1) represents the source area of the Kolyma River. No dams exist in this subbasin of 10,300 km^2. The seasonal cycle shows the highest flow in June (405 $m^3 s^{-1}$) and the lowest flow in April (1 $m^3 s^{-1}$) (Fig. 2.2a). The start of snowmelt discharge is in May (95 $m^3 s^{-1}$); the other high flow months are July (240 $m^3 s^{-1}$), August (189 $m^3 s^{-1}$), and September (139 $m^3 s^{-1}$). Low flows from November to April range from 42 to 1 $m^3 s^{-1}$ (Fig. 2.2a). The interannual variation of monthly discharge, measured by standard deviation, is small from October to April (17–1 $m^3 s^{-1}$), and large during June to August (199–78 $m^3 s^{-1}$) (Fig. 2.2b). Flow data during 1942–94 show (total) monthly streamflow increased strongly by 80 (53%) and 33 $m^3 s^{-1}$ (27%) in August and September, respectively, and weakly by 2 $m^3 s^{-1}$ (17%) in November and 38 $m^3 s^{-1}$ (49%) in May (Fig. 2.2c). On the other hand, discharge changed very little in January, and decreased by 25 $m^3 s^{-1}$ (−6%) in June and by 27 $m^3 s^{-1}$ (−11%) in July (Fig. 2.2c). The annual discharge increased by 10% during 1942–94 (Fig. 2.3).

b. Orotuk (upper basin)

The Orotuk station (B in Fig. 2.1) is located in the upper Kolyma basin; it has a drainage area of 43,600 km^2. Flow pattern at the Orotuk station is similar to that of the Kulu station, with high flow ranging from 360 to 81 m^3s^{-1} during May to October, and low flow from 25 to 2 m^3s^{-1} during November to April (Fig. 2.2a). This flow pattern is typical for regions with continuous permafrost. Because of limited subsurface storage in the shallow active layers, it is typical to have very low flow in the winter and very high peak flow in summer. The ratio between June (highest) and April (lowest) flows is about 500 in the Orotuk sub basin. Standard deviations of monthly flows are low from November to April (1-9 m^3s^{-1}), and high during May to August (243-280 m^3s^{-1}), with the maximum in June (539 m^3s^{-1}) (Fig. 2.2b). Discharge changes during 1957-97 over the Orotuk sub basin are characterized by positive trends for all the months. Total increase during 1957-97 are weak (2-6 m^3s^{-1}) over the cold season (November to April) and strong in May (195 m^3s^{-1}), June (184 m^3s^{-1}), July (183 m^3s^{-1}), August (177 m^3s^{-1}), and September (186 m^3s^{-1}) (Fig. 2.2c). Streamflow changes in this unregulated sub basin are possibly due to climatic changes. The annual discharge shows a strong increase of 22% during 1957-97 (Fig. 2.3).

c. Duscania (upper basin)

The Duscania station (C in Fig. 2.1) is located in the upper Kolyma basin; it has a drainage area of 50-100 km^2. Duscania shows similar pattern to that of the Kulu station, with high flow ranging from 364 to 92 m^3s^{-1} from May to October, and low flows from 26 to 2 m^3s^{-1} during November to April (Fig. 2.2a). Standard deviations of monthly flows are low from November to April (2-11 m^3s^{-1}), and high during May to August (351-268 m^3s^{-1}), with the maximum in June (809 m^3s^{-1}) (Fig. 2.2b). Discharge changes during 1948-80 at the Duscania are characterized by negative (total) trends in June (79 m^3s^{-1}) and November (2 m^3s^{-1}). May has a positive trend of 261 m^3s^{-1}, while low flow months during December to April show very weak positive trends of less than 2 m^3s^{-1} (Fig. 2.2c). The annual discharge increased by 19% during 1948-80 (Fig. 2.3).

d. Sinegor'e (upper basin)

The Sinegor'e station (D in Fig. 2.1) is situated in the upper Kolyma basin. This station controls a drainage area of 61,500 km^2. Flow data are available from 1932 to 1989, with missing data during 1952-67 (a gap of 15 yr). The seasonal cycle of monthly streamflow shows high discharge (160- 532 m^3s^{-1}) from May to October, with the maximum discharge in June (1762 m^3s^{-1}), and very low flows from November (54 m^3s^{-1}) to April (10 m^3s^{-1}) (Fig. 2.2a). A reservoir was built in the main river valley above the Sinegor'e station in the early 1980s. The dam is 130 m high, 780 m long, and its volume is 10 km^3. It is a rock-filled dam (with a soil core) with a surface area of 441,103 km^2 (Petrov and Losev 1976). The Kolymskoye is the biggest dam in the Kolyma River basin and the reservoir was filled during 1986-90. A power plant (capacity of 3.26 X 109 kWh^{-1}) was also built below the dam. Most of the reservoirs in Siberia were designed for multipurpose use, such as water supply and hydropower generation. To quantify the dam impact on streamflow regime and change, we compare the mean monthly flows between the pre- and post-dam periods. The results show discharge decreases during the post-dam period by 498-397 m^3s^{-1} (or about 85%-36%) during May to July, and increases by 136-134 m^3s^{-1} (525%-1,151%) from December to January (Fig. 2.4a). Monthly discharge records clearly show sudden changes in streamflow right after the dam completion due to reservoir regulation (Fig. 2.5). Streamflow increases in cold season (November- April) because reservoir releases water to generate more power in winter season; discharge decreases in June and July are due to reservoir storing water to reduce the snowmelt and rainfall floods (Fig. 2.5). The monthly flow records also suggest that, as result of dam regulation, monthly discharge varied within the range of natural variability (the difference between the maximum and minimum flow during the pre-dam period) over the high flow season, and fluctuated out of the range of natural variability during the low flow season due to significant increases in discharge over the post dam period (Fig. 2.5). Similar discharge changes have been found in other regulated Siberian watersheds, such as the Lena, Yenisei, and Ob basins (Ye et al. 2003; Yang et al. 2004a). Ye et al. (2003) also reported that dam regulation affects long-term streamflow trends in Siberia. Their study reported overestimation of trends in winter and underestimation of trends in summer. Because of many missing discharge data at this station, dam impacts on streamflow trends cannot be determined.

e. Ust'-Srednekan (upper basin)

The Ust'-Srednekan station (E in Fig. 2.1) controls a drainage area of 99,400 km^2. Similar to the upper basin, monthly discharge at the Ust'-Srednekan station is low from November (118 m^3s^{-1}) to April (52 m^3s^{-1}) and high during May to July (969–1,621 m^3s^{-1}), with the peak flow in June (2,945 m^3s^{-1}) (Fig. 2.2a). The inter-annual variation follows the pattern of monthly flow, with the maximum in June (1,586 m^3s^{-1}) and minimum in April (100 m^3s^{-1}) (Fig. 2.2b). The standard deviations are small (100-108 m^3s^{-1}) for the low flow months from November to April, and large (600-687 m^3s^{-1}) for the high flow months during May to July (Fig. 2.2b). The Ust'-Srednekan station is located 1,493 km downstream of the dam. Usually dam impact is most significant at the location right below the reservoir. It is important, however, to note that the influence of the reservoir is clearly seen at the Ust'-Srednekan station. For example, comparisons of mean flows between the pre- and post-dam periods show flow increases for most months, except during May to July (Fig. 2 . 4b). Strong streamflow increase during the post-dam period by about 205 m^3s^{-1} (or 522%-3,157%) in cold season (December-April) and decrease in June peak flow by 133 m^3s^{-1} (41%) are the most typical indications of dam regulation. Furthermore, long-term (total) trend during 1933-2000 also show moderate decreases from May (70 m^3s^{-1}) to July (15 m^3s^{-1}), and significant (at 99% confidence) increases during November (209 m^3s^{-1}) to April (259 m^3s^{-1}). These changes in seasonal streamflow characteristics are mainly due to reservoir recharge during the summer and fall seasons (Ye et al. 2003). As the result of monthly flow changes, the annual flow has increased by 2% during 1933-2000 (Fig. 2.3).

f. Yasachnaya at Nelemnoy (unregulated tributary/middle basin)

The Yasachnaya station (F in Fig. 2.1) is located on a western tributary in the middle Kolyma basin. It has a drainage area of 32,000 km^2 and contributes 6% of the total Kolyma discharge. There are no dams in this sub basin. Flow data are available from 1977 to 1988. Monthly mean flow is typical, with highest discharge in June (1,367 m^3s^{-1}) and lowest in April (18 m^3s^{-1}). The standard deviations show the same variation as monthly mean flows. The monthly discharge records in this unregulated sub basin did not show significant increases for the cold season (November- April) and very slight changes during the high flow months (May-July) (Fig. 2.5). The annual flow decreased by about 10% during the 12 yr (1977-88) (Fig. 2.3).

g. Srednekolunsk (lower basin)

The Srednekolunsk station (G in Fig. 2.1) is located in the lower part of main Kolyma River valley, 1,720 km downstream of the reservoir. This station controls a drainage area of

361,000 km^2. Monthly mean discharge, standard deviation, and trend generally show similar patterns with other upstream stations, such as low flows (115-241 m^3s^{-1}) during April to December and the peak flow (2,107 m^3s^{-1}) in June (Fig. 2.2a). Reservoir regulation alters the seasonal discharge pattern (Shiklomanov 2000; Yang et al. 2003). The dam effect is still visible at this station, although the impact is much weaker relative to the upstream stations near the dam. Comparisons of pre- and post-dam mean monthly flows (Fig. 2.4c) show low flow increases by 130 (43%) to 268 m^3s^{-1} (454%) during November to April, and high discharge decreases by 2,550 to 519 m^3s^{-1} during June to August in the post-dam era (1988-2000) (Fig. 2.2b). Monthly discharge records (Fig. 2.5) show long term increases during November to April and slight decreases during May to July. The trend analyses of monthly flow data during 1927-2000 show significant increases in the cold season. The most extreme trend is in January, with the total increase of 338 m^3s^{-1} (4,110%), and followed by February, 337 m^3s^{-1} (1,234%), March, 323 m^3s^{-1} (868%), and April, 300 m^3s^{-1} (781%). The high flow season and early fall showed decreasing trends of 2,701 m^3s^{-1} (30%) in June, 1703 m^3s^{-1} (4%) in July, 584 m^3s^{-1} (3%) in August, and 108 m^3s^{-1} (5%) in September, respectively (Fig. 2.2c). The t-test results reveal high significance for the low flow months, November, December, January, February, and April being statistically significant at 98%-99% confidence. Trends in May, June, and July flows are significant at 60%, 95%, and 98% confidence, respectively. Annual discharge has a decreasing tendency (-11%) over the 72 yr (1927–2000) (Fig. 2.3). The effect of dam regulation is strong and easy to detect during the low flow season than the high flow season. To examine how the impact of dam regulation has been transferred downstream, we analyzed discharge budget along a selected section of the main river valley. We calculated the mean monthly discharge during 1978-2000 between the pre- and post-dam periods for the Srednekolunsk and the Ust'-Srednekan stations. The results show that low flow changes during November to April are about 2.3 km^3 at the Ust'-Srednekan station (near the dam site) and 3.5 km^3 at the Srednekolunsk station. The regulation effect is expected to decrease from the dam site to downstream. However, flow increases are higher at the Srednekolunsk station and lower at the dam site. The Srednekolunsk station is 1,720 km downstream of the dam. It receives runoff contribution from other tributaries. Yearly flow at the Srednekolunsk station is much higher (71 km^3) than that (25 km^3) near the dam site. Because of lack of flow data for the tributaries above the Srednekolunsk station, it is impossible to use discharge data at these stations (currently available for this analysis) to accurately determine the impact of dam regulation in the mid- and lower parts of the basin. Additional flow data are necessary to calculate the tributary runoff contribution and secular streamflow budget. Differences in both yearly flow amount and trend found between the Srednekolunsk station and the dam site seem to imply that dam regulation cannot account for 100% of the observed increases in winter discharge near the basin outlet. It also suggests that other factors, such as climate effects and permafrost changes, may influence streamflow change over this part of the basin.

h. Kolymskoye (lower basin)

The Kolymskoye station (station H in Fig. 2.1) is located above the mouth of the Kolyma River; it is the nearest station to the basin outlet. Discharge data collected at or near the basin outlet are particularly important, as they have been used for validations of hydrological/land surface models and GCMs. The monthly flow data at the Kolymskoye station are available during 1978-2000. The monthly mean discharge (Fig. 2.2a) is low from November to April (47-18 m^3s^{-1}) and high during May to June (178-754 m^3s^{-1}), with the highest flow in June (1490 m^3s^{-1}) during the snowmelt season. Discharge from August to October varies from 602 to 180 m^3s^{-1}. The peak flow in June is about 80 times greater than the lowest flow in April. Following the pattern of monthly mean discharge, the standard deviations of monthly flow are low for November to April (129-109 m^3s^{-1}) and higher during May to June (1464-4844 m^3s^{-1}) (Fig. 2b). Comparisons of the long-term mean streamflow between the pre- and post-dam periods demonstrate significant changes (Fig. 2.4d). In the post-dam era, peak flows decreased by 217 m^3s^{-1} (1%) in June, 3,710 m^3s^{-1} (37%) in July, and 1,189 m^3s^{-1} (17%) in August. Discharge increased by 1,051 m^3s^{-1} in September and by 433 m^3s^{-1} in October, respectively. The most significant impact of reservoir regulation is seen during the cold season from November to April, when flows increased by 32% (127 m^3s^{-1}) in December to 208% (164 m^3s^{-1}) in April. The Kolymskoye station is located 1902 km downstream of the reservoir. Streamflow routing and runoff contribution from tributaries subdue the effect of dam regulation. Monthly flow records during 1978-2000 at this station (Fig. 2 . 5) show significant changes. Total trends over 1978-2000 were 229 to 250 m^3s^{-1} during November to April (Fig. 2.2c), with the highest increase of 250 m^3s^{-1} (400%) in April. Changes in March and February were 275 (350%) and 263 m^3s^{-1} (275%), respectively. Negative trends were found for July, 4982 m^3s^{-1} (-49%), and August, 1490 m^3s^{-1} (-22%) (Fig. 2.2c), and a weaker increase of 925 m^3s^{-1} (6%) was seen in June. The remaining months have positive trends of 40%-109%. Annual discharge at the Kolymskoye station shows a weak decrease of 1.5% during 1978-2000 (Fig. 2.3). This decrease seems to be consistent with climate changes over the basin, since basin yearly temperature increased and precipitation decreased slightly during 1978-2000 (Fig. 2 . 6). Dam regulation also affects the inter-annual variation in streamflow. To determine the changes in flow variation, we calculated and compared the standard deviation of monthly streamflow between the pre- and post- dam periods at the four stations downstream of the dam (Fig. 2.7). The results show that, relative to the pre-dam period, the post-dam streamflow variations increase during the low flow months (November-April), and decrease in high flow season (June-July). This means that dam impacts on flow variation differ over the season.

i. Eastern tributaries

There are five major unregulated tributaries in the eastern part of the Kolyma basin. To understand the streamflow characteristics and changes mainly due to natural causes over the eastern Kolyma basin, two control stations on the main eastern tributaries, that is, the Bol'shoy Anuy (station I in Fig. 2.1) at the Konstantinovo Valley and the Oloy station (J in Fig. 2.1) at the Utuchan Valley were analyzed in this study. The Bol'shoy Anuy station has a drainage area of 49,600 km^2 and the Oloy station controls a drainage area of 15,700 km^2. The monthly mean flows at the Bol'shoy Anuy station are characteristic of the Kolyma basin, with high discharge (292-67 m^3s^{-1}) from May to October, the maximum discharge in June (1272 m^3s^{-1}), and very low flows from November (16 m^3s^{-1}) to April (2 m^3s^{-1}) (Fig. 2.8a). Flow records collected at the Oloy station during 1975-98 show that monthly mean streamflow are high (40-35 m^3s^{-1}) from May to October, with the maximum discharge in June (488 m^3s^{-1}), and very low discharge from November (13 m^3s^{-1}) to April (7 m^3s^{-1}) (Fig. 2.8a). Trend analyses for the Bol'shoy Anuy station during 1978-2000 show total flow decreases in May (-17 m^3s^{-1}), July (-426 m^3s^{-1}), August (-135 m^3s^{-1}), and October (-20 m^3s^{-1}); positive trends in June (239 m^3s^{-1}), September (82 m^3s^{-1}), and little changes during November to April (Fig. 2.8b). These monthly flow changes caused an annual discharge decrease of 8% during 1978-2000. Discharge data records at the Oloy station during 1975-88 indicate increasing trends (0.2-9.8 m^3s^{-1}) during November to April, and decreasing trends of 0.8 m^3s^{-1} in June, 33.9 m^3s^{-1} in July, and 85.5 $m^3s^{-1}$for September. Annual discharge decreased by 3 m^3s^{-1} (3%) over the period 1975-88. It is important to notice the clear difference in flow trends between the regulated and unregulated sub basins of the Kolyma River. For instance, low flow trends are much smaller in the unregulated eastern tributaries (Fig. 2.8b) and very high in the regulated main valley, such as at the Kolymskoye station near the basin outlet (Fig. 2.2a). This clearly demonstrates the effect of dam regulation that caused winter flow increases in the main river valley.

2.5 Conclusions

Climate over the Arctic regions has significantly changed in the last decades (Chapman and Walsh 1993; Serreze et al. 2000). This study investigates Kolyma River hydrologic regimes and changes induced by reservoir regulation and climate variations. Streamflow records show that the Kolyma basin has the basic characteristic of permafrost regions, with low flow from November to April and high discharge during May to August. Changes in monthly discharge are different for the upper, middle, and lower parts of the basin. In the upper basin without dam regulation, streamflow increased for most months. The increases were weak during November to May, and strong for August and September, while flow decreased weakly in June and July over the Kulu Valley (source of the Kolyma basin). The two eastern tributaries did not show consistency in streamflow trends. In the Konstantine Valley, discharge decreased weakly in March, May, and October and strongly in July and August; and increased strongly in both June and September and weakly during November–February and in April.

Dams have a major influence on watershed storage, discharge regime, and change (Vörösmarty et al. 1997; Yang et al. 2004 a, b; Ye et al. 2003). Over the mid– lower parts of the basin (downstream of the dam), streamflow increased during the low flow season, and decreased in the high flow months, because reservoirs store water during the peak flow season and release water during the low flow season. For instance, over the post-dam period (1986- 2000), streamflow at the Ust'-Srednekan station (1,493 km downstream of the dam) suddenly increased by 522%-3,157% during December to April, and decreased by 41% in July. Similarly, comparisons of pre- and post-dam mean monthly flows at the Srednekolunsk station (1,720 km downstream of the dam) show low flow increases of 130 m^3s^{-1} (43%) - 268 m^3s^{-1} (454%) from November to April, and high discharge decreases by 2,550-519 m^3s^{-1} during June to August in the post-dam era. Comparisons of flow changes between the two downstream stations demonstrate the diminishing effect of the dam as the distance between the dam and the station increases. Changes in seasonal streamflow, particularly the increases in low flows, over the mid–lower reaches of the Kolyma basin is partly due to dam regulation. A study by Ye et al. (2003) for the Lena basin found similar results, and they reported that cold season increase in discharge is a combination of both reservoir regulation and natural runoff changes. Because of the dramatic hydrologic changes caused by the dam regulation in the Kolyma watershed, it is difficult to detect streamflow changes due to natural causes in the regulated mid–lower basins.

Annual discharge records show different changes within the Kolyma basin. The upper basin (upstream of the dam) shows yearly runoff increases of 10%, 19%, and 22% at the Kulu, Duscania, and Orotuk stations, respectively. The two eastern tributaries have flow decreases by 8% and 3% at the Bolsoy Anuy and Oloy stations, respectively. Annual flow decreased by about 2%-4% over the mid-lower basin (downstream of the dam) partly due to reservoir regulation. Similar to the temperature changes reported in Symon et al. (2005) over the Arctic regions, the Kolyma basin climate records show an annual warming trend of 0.4°C during 1930-2000. The seasonal basin temperatures show a cooling trend during November to December. Similar changes in winter temperatures were reported by Chapman and Walsh

(1993) for the east Siberian regions. Basin yearly precipitation showed a decrease during the last 4-5 decades. These changes are somewhat consistent with river runoff decrease over the Kolyma basin as a whole; discharge changes in the unregulated sub-basins are perhaps mainly due to climate variations. However, given the weak changes in climate over the basin, the weak decreases in yearly runoff near the basin outlet is also likely due to dam impact, such as in- filtration and evaporation water losses from the large reservoir (Jansen 1988). Regardless of the cause, it is important to note that the Kolyma River discharge into the Arctic Ocean has decreased slightly over the past 20 yr. Long-term (1927-2000) data collected at the Srednekolunsk station (181 km above the Kolymskoye station) show discharge decrease by 11%. The Kolyma basin is one of the six largest Siberian rivers analyzed by Peterson et al. (2002), and this river showed no increase in yearly discharge.

Seasonal discharge changes are stronger and easier to detect than the annual flow changes. Recent analyses have shown discharge increases in the winter season for the Ob, Yenisei, and Lena Rivers (Ye et al. 2003; Yang et al. 2004 a, b), where human impacts are predominant due to dam constructions, farming, mining, and other activities. Macdonald et al (2002) suggested that timing of freshwater input is important and change in seasonal streamflow, increase in winter flow at the expense of summer inflow, could stall convection of shelf. On the other hand, insignificant change was found for the less developed regions, such as the Mackenzie basin in northern Canada (Dynesius and Nilsson 1994; Revenga et al. 1998; Déry and Wood 2005). This study identifies major changes in streamflow seasonal cycle over the Kolyma basin due to reservoir regulation; it clearly underlines the importance of human activities in regional and global environment changes. Further research is necessary to better understand climate impact on river discharge changes, particularly the relationships among river streamflow, basin temperature, precipitation, and snow cover changes.

2.6 Acknowledgments

This study was sponsored by NSF Grants OPP-0230083 and ARC-0612334, and NOAA/ CIFAR Grant NA17RJ1224.

2.7 Figures

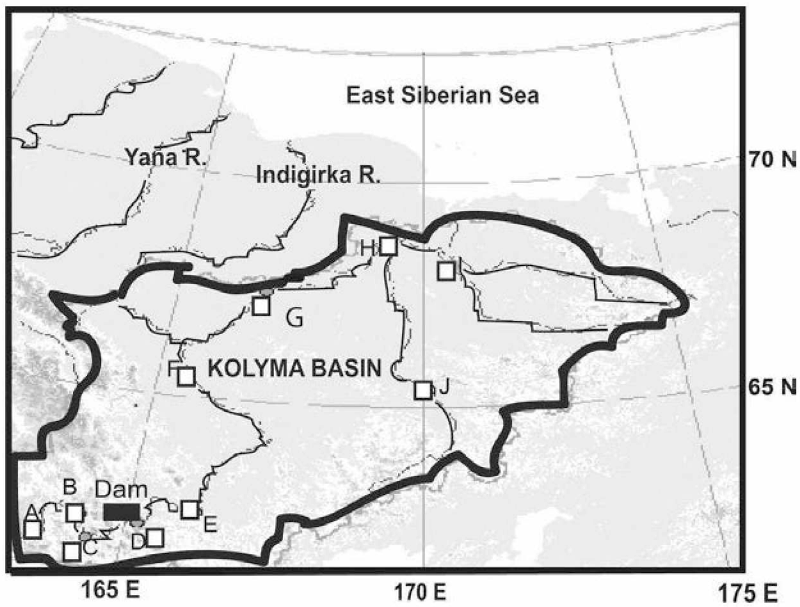

Figure 2.1. The Kolyma watershed and the location of the dam and hydrologic stations used in this study.

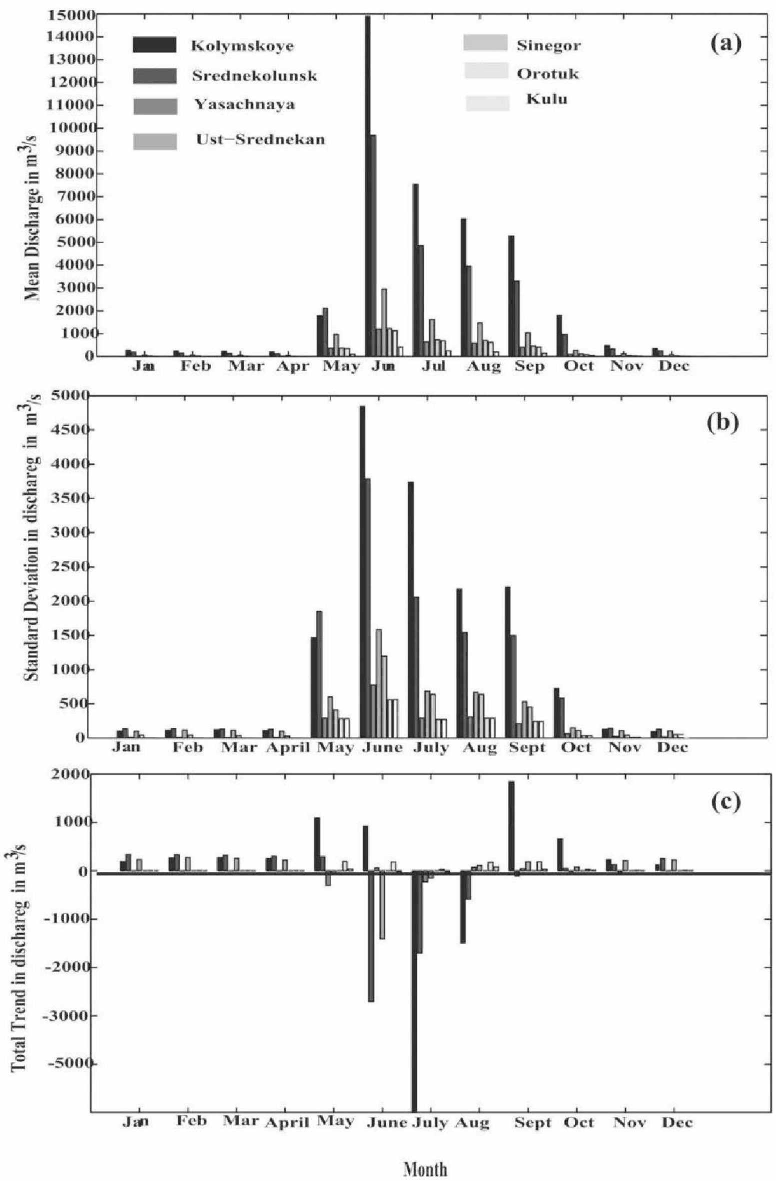

Figure 2.2. (a) Monthly mean discharge, (b) standard deviation of discharge, and (c) trend analysis for selected stations on the Kolyma River. (Refer to Table 1 for data period for every station.)

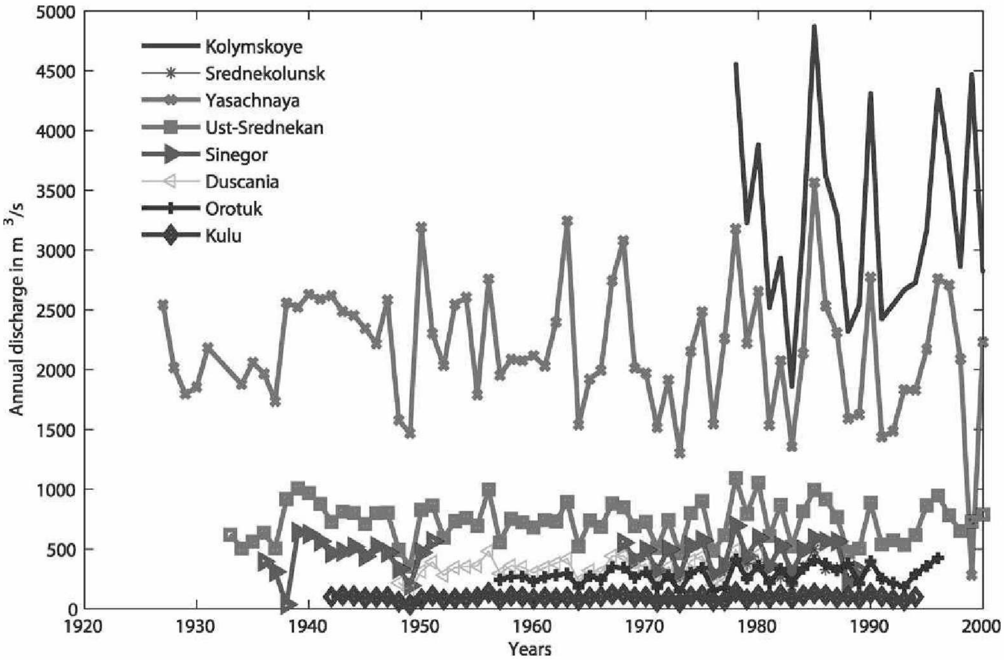

Figure 2.3. Annual discharge plots for selected hydrologic stations in the basin.

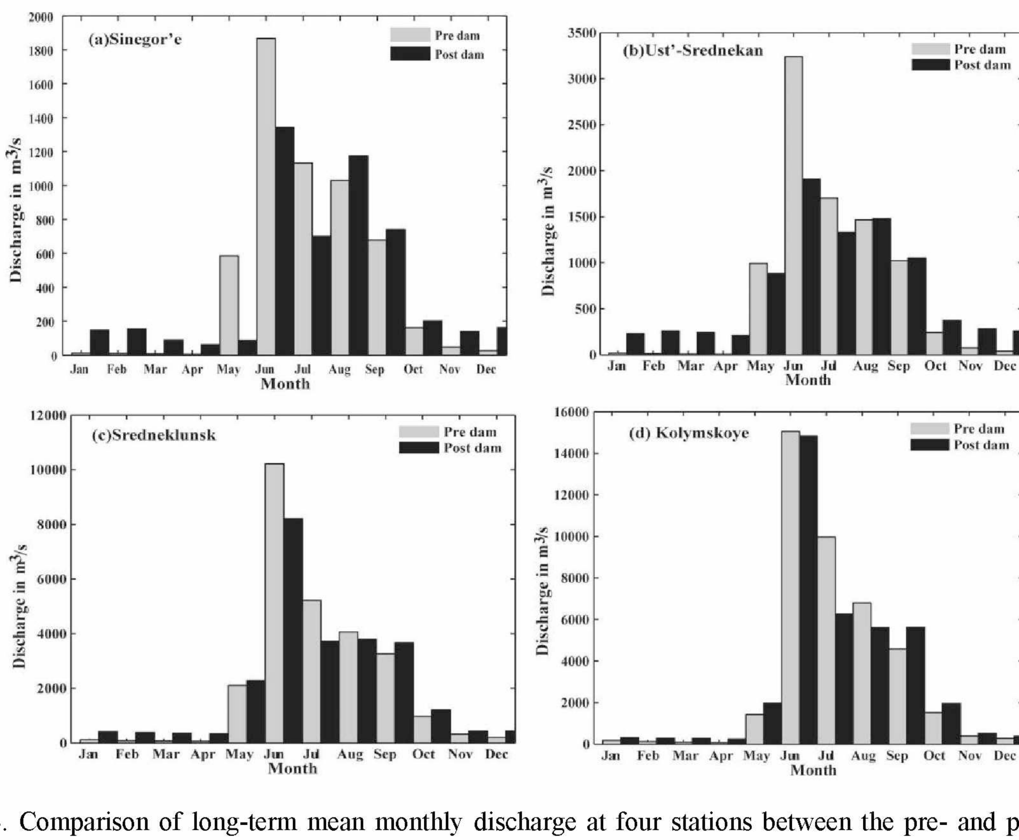

Figure 2.4. Comparison of long-term mean monthly discharge at four stations between the pre- and post-dam periods.

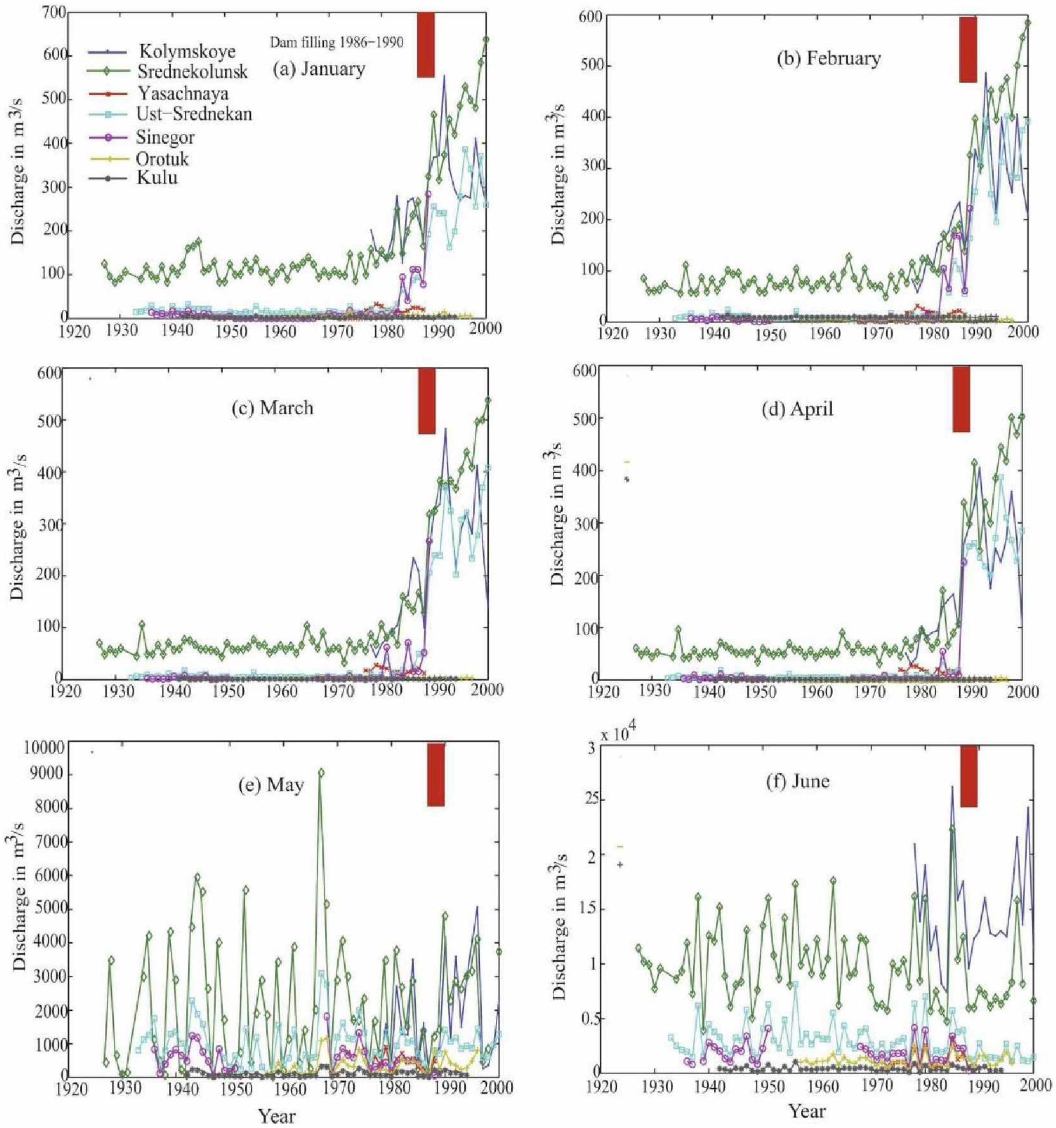

Figure 2.5. Monthly discharge records at selected hydrologic stations in the basin; the bar indicates the filling period from 1986 to 1990.

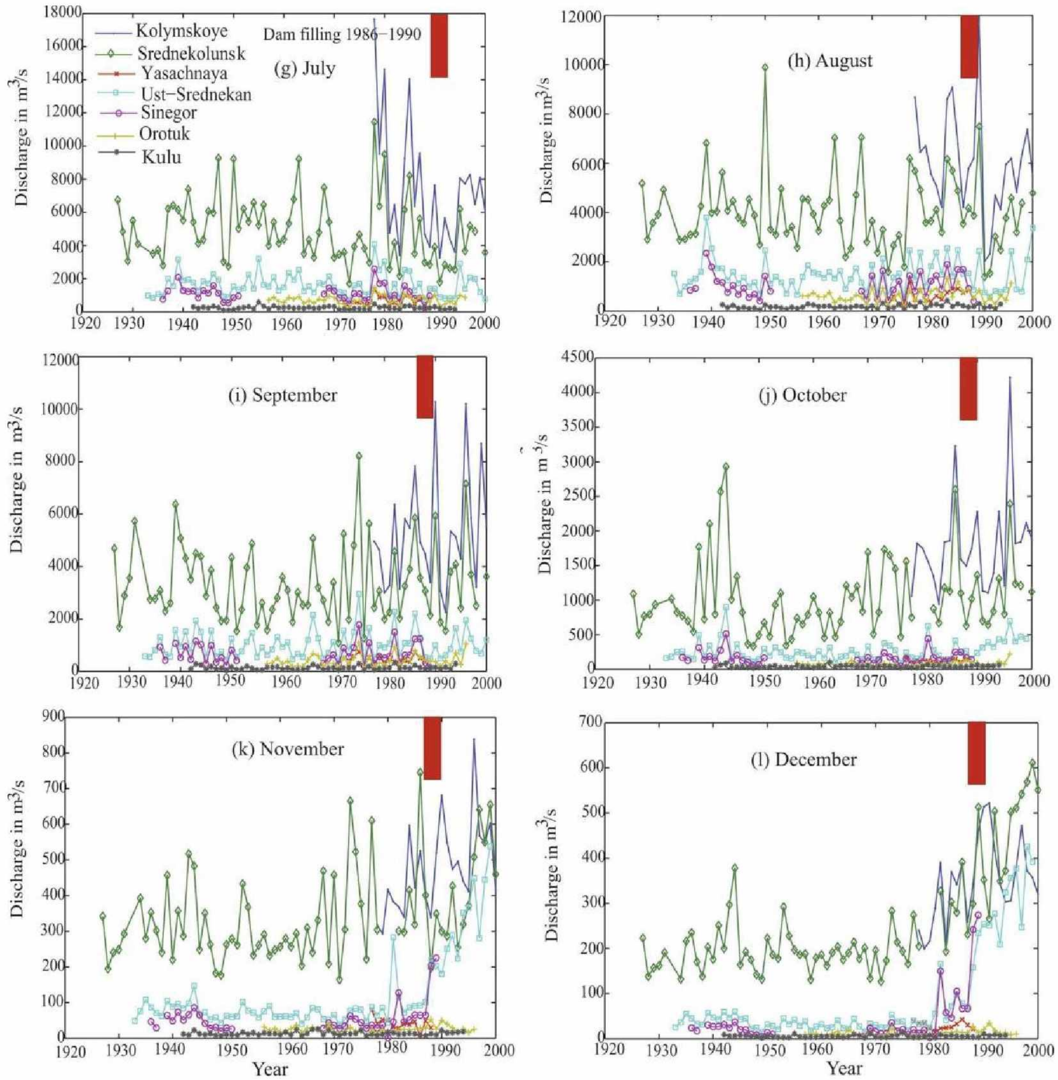

Figure 2.5. (*Continued*)

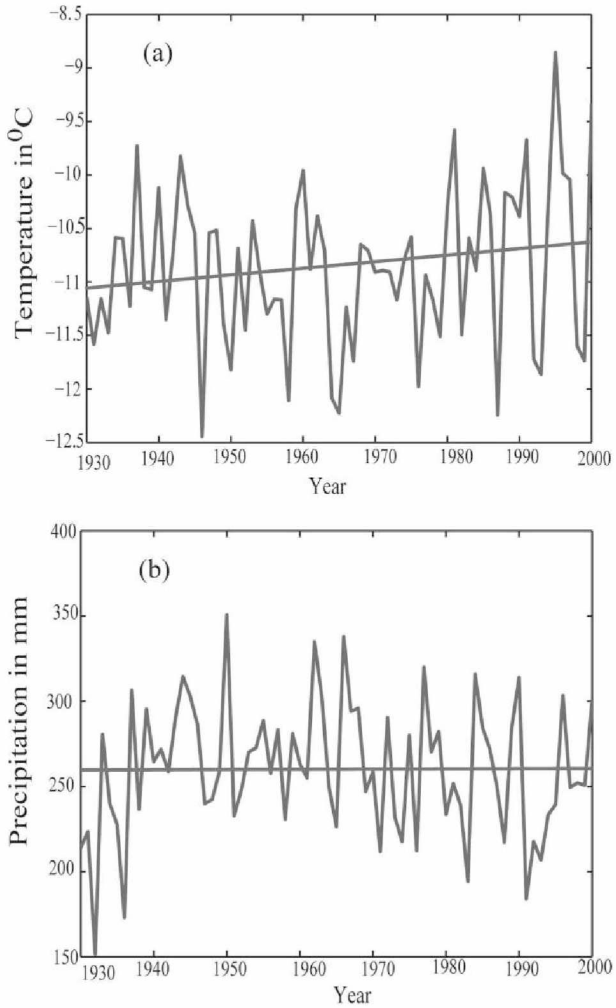

Figure 2.6. (a) Basin-mean annual temperature and (b) annual precipitation records from 1930-2000.

2.8 Tables

Table 2.1. List of hydrologic stations used in this study.

Station code (Fig. 2.1)	Station name/location	Latitude (°E)	Latitude (°N)	Data period	Discharge area x 1000 km^2	% of Kolyma Basin	Annual runoff (km^3)	% of basin runoff	Dam information
H	Kolymskoye/basin outlet	68.73	158.72	1978–2000	526	100	103	100	Downstream of dam
G	Srednekolunsk/main river valley	67.47	153.69	1927–2000	361	69	68	67	Downstream of dam
F	Yasachnaya/tributary	65.4	151.07	1972–88	32	6	9	9	Unregulated subbasin
E	Ust'-Srednekan /main river valley	62.43	152.3	1933–2000	99	19	23	22	Downstream of dam
D	Sinegor's/main river valley	62.07	150.47	1936–89	61	12	10	9	Downstream of dam
C	Duscania/main river valley	61.65	148.83	1948–80	50	10	11	10	Upstream of dam
B	Orotuk/main river valley	62.12	148.47	1957–97	43	8	9	8	Upstream of dam
A	Kulu /main river valley	61.9	147.42	1942–94	10	2	3	3	Upstream of dam
I	Bol'shoy Anuy/ Konstantinovo	68.15	161.16	1978–2000	49	12	8	8	Unregulated subbasin
J	Oloy /Utuchan	65.67	162.43	1975–88	15	5	4	3	Unregulated subbasin

2.9 References

Aagaard, K., and E. C. Carmack, 1989: The role of sea ice and other freshwater in the Arctic circulation. *J. Geophys. Res.*, **94,** 14 485–14 498.

Berezovskaya, S., D. Yang, and D. L. Kane, 2004: Compatibility analysis of precipitation and runoff trends over the larger Siberian watersheds. *Geophys. Res. Lett.,* **31,** L21502, doi:10.1029/2004GL021277.

Chapman, W. L., and J. E. Walsh, 1993: Recent variations of sea ice and air temperature in high latitudes. *Bull. Amer. Meteor. Soc.,* **74,** 33–47.

Déry, S. J., and E. F. Wood, 2005: Decreasing river discharge in northern Canada. *Geophys. Res. Lett.,* **32,** L10401, doi:10.1029/2005GL022845.

Dynesius, M., and C. Nilsson, 1994: Fragmentation and flow regulation of river systems in the northern third of the world. *Science,* **266,** 753–762.

Foley, J. A., J. E. Kutzbach, M. T. Cole, and S. Levis, 1994: Feedbacks between climate and boreal forests during Holocene Epoch. *Nature,* **371,** 52–54.

Hinzman, L. D., and Coauthors, 2005: Evidence and implications of recent climate change in northern Alaska and other arctic regions. *Climate Change*, **72**, 251–298.

Holmes, R. M., J. W. McClelland, and B. J. Peterson, 2003: Consideration of permafrost thaw as a significant contributor to increasing Eurasian Arctic river discharge. *Extended Abstracts, SEARCH Open Science Meeting,* Seattle, WA, National Science Foundation's Office of Polar Programs, 27–29.

Hulme, M., 1991: An intercomparison of model and observed global precipitation climatologies. *Geophys. Res. Lett*., 18, 1715-1718.

IUCN, 2006: Water resources eAtlas. [Available online at http://www.iucn.org/themes/wani/eatlas/index.html.]

Jansen, R. B., 1988: Advanced Dam Engineering for Design, Construction, and Rehabilitation, *Springer*, 811 pp.

Jones, P. D., 1994: Hemispheric surface air temperature variations: A reanalysis and an update to 1993. *J. Climate,* **7,** 1794–1802.

Kane, D. L., 1997: The impact of Arctic hydrologic perturbations on Arctic ecosystems induced by climate change. *Ecol. Stud.,* **124,** 63–81.

Kane, D. L., R. E. Gieck, and L. D. Hinzman, 1990: Evapotranspiration from a small Arctic watershed. *Nord. Hydrol.*, **21,** 253–272.

Kuchment, L. S., A. N. Ge'fan, and V. N. Demidov, 2000: A model of runoff formation on watersheds in the permafrost zone: Case study of the Upper Kolyma River. *Water Resour. Res.*, **27,** 392–400.

Lammers, R. B., A. I. Shiklomanov, C. J. Vörösmarty, B. M. Fekete, and B. J. Peterson, 2001: Assessment of contemporary Arctic river runoff based on observational discharge records. J. Geophys. Res., **106**, 3321–3334.

Macdonald, R. W., F. A. McLaughlin, and C. E. Carmack, 2002: Freshwater and its sources during the SHEBA drift in the Canada Basin of the Arctic Ocean. *Deep-Sea Res. I,* **49,** 1769– 1785.

McClelland, J. W., R. M. Holmes, B. J. Peterson, and M. Stieglitz, 2004: Increasing river discharge in the Eurasian Arctic: Consideration of dams, permafrost thaw, and fires as potential agents of change. *J. Geophys. Res.*, **109**, D18102, doi:10.1029/2004JD004583.

Miah, M. A., 2002: Man-made climatic changes in the Ganges basin. *Int. J. Climatol.,* **22,** 993– 1016.

Nelson, F. E., and O. A. Anisimov, 1993: Permafrost zonation in Russia under anthropogenic climate change. *Permafrost Peri-glacial Processes,* **4** (2), 137–148.

Oechel, W. C., S. J. Hastings, G. Vourlitis, M. Jenkins, G. Richers, and N. Guilke, 1993: Recent changes of Arctic tundra eco systems from a net carbon dioxide sink to a source. Nature, 361, 520–523.

Panfilova, V. K., 1986: The water balance of the upper Kolyma basin. *Proc. Cold Region Hydrology Symp.,* Fairbanks, AK, American Water Resources Association, 293–296.

Peterson, B. J., R. M. Holmes, J. W. McClelland, C. J. Voros-marty, R. B. Lammers, A. I. Shiklomanov, I. A. Shiklo-manov, and S. Rahmstorf, 2002: Increasing river discharge to the Arctic Ocean. *Science,* **298,** 2171–2173.

Petrov, V. G., and E. D. Losev, 1976: Dam of the Kolyma hydro-electric power plant. Cold Regions Research and Engineering Laboratory Rep. 563, 12 pp.

Revenga, C., S. Murray, J. Abramovitz, and A. Hammond, 1998: *Watersheds of the World: Ecological Value and Vulnerability*. Water Resources Institute, Worldwide Institute, 172 pp.

Serreze, M. C. 2003. Arctic Climate. In *Encyclopedia of Atmospheric Sciences*, ed. J. R. Holton, J. A. Curry and J. A. Pyle. Academic Press.

Serreze, M. C., and Coauthors, 2000: Observation evidence of recent change in the northern high-latitude environment. *Climatic Change,* **46,** 159–207.

Serreze, M. C., D. H. Bromwich, M. P. Clark, A. J. Etringer, T. Zhang, and R. B. Lammers, 2003: Large- scale hydro-climatology of the terrestrial Arctic drainage system. *J. Geophys. Res.,* **108,** 8160, doi:10.1029/2001JD000919.

Shiklomanov, I. A., 2000: Appraisal and assessment of world water resources. *Water Int.,* **25,** 11–32.

Shiklomanov, I. A., and A. I. Shiklomanov, 2003: Climatic change and the dynamics of river runoff into the Arctic Ocean. *Water Resour. Res.,* **30,** 593–601.

Swanson, D. K., 2003: A comparison of taiga flora in north-eastern Russia and Alaska/Yukon. *J. Biogeogr.,* **30,** 1109–1121.

Symon, C., and Coauthors, 2005: *Arctic Climate Impact Assessment.* Cambridge University Press, 1042 pp.

Tsuyuzaki, S., T. Ishizaki, and T. Sato, 1999: Vegetation structure in gullies developed by the melting of ice wedges along Kolyma River, northern Siberia. Ecol. Res., 14, 385–391.

Vörösmarty, C. J., K. Sharma, B. Fekete, A. H. Copeland, J. Holdon, J. Marble, and J. A. Lough, 1997: The storage and aging of continental runoff in large reservoir systems of the world. Ambio, 26, 210–219.

Vörösmarty, C. J., B. Fekete, M. Meybeck, and R. B. Lammers, 2000: The global system of rivers: Its role in organizing continental land mass and defining land-to-ocean linkages. *Global Bio- geochem. Cycles,* **26,** 210–219.

Walsh, J. E., 2000: Global atmospheric circulation patterns and relationships to Arctic freshwater fluxes. *The Freshwater Budget of the Arctic Ocean,* E. Lewis et al., Eds., NATO Science Partnership Sub-Series 2: Environmental Security, Vol. 70, Kluwer Academic, 21–41.

Welp, L. R., J. T. Randerson, J. C. Finlay, S. P. Davydon, G. M. Zimova, A. I. Davydova, and S. A. Zimov, 2005: A high-resolution time series of oxygen isotopes from the Kolyma River: Implications for the seasonal dynamics of discharge and basin-scale water use. Geophys. Res. Lett., 32, L14401, doi:10.1029/2005GL022857.

Yang, D., D. Robinson, Y. Zhao, T. Estilow, and B. Ye, 2003: Streamflow response to seasonal snow cover extent changes in large Siberian watersheds. *J. Geophys. Res.*, **108,** 4578, doi:10.1029/2002JD003149.

Yang, D., B. Ye, and A. Shiklomanov, 2004: Discharge characteristics and changes over the Ob River watershed in Siberia. *J. Hydrometeor.*, **5,** 595–610.

Yang, D., B. Ye, and D. Kane, 2004b: Streamflow changes over Siberian Yenisei river basin. *J. Hydrol.*, **296,** 59–80.

Ye, B., D. Yang, and D. L. Kane, 2003: Changes in Lena River streamflow hydrology: Human impacts versus natural variations. *Water Resour. Res.*, **39,** 1200, doi:10.1029/2003WR001991.

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3. Streamflow analysis for the Yana basin in eastern Siberia[2]

3.1 Abstract

We analyze Yana River streamflow and climate data in order to understand climate change and its impact on basin hydrology. Basin temperature and precipitation records show little change during 1977-1999. Discharge data near the basin mouth suggest changes (increase and decrease) over the summer months. Basin precipitation has a positive correlation with discharge during June, July and August. The relationship between snow water equivalent and discharge follows an inverse relation; maximum snow water equivalent and discharge have a linear relation, with inconsistencies in some years. Further examination is needed to improve this relationship. The results of this study are useful for a better understanding of the hydrological regime and changes over the northern regions.

3.2 Introduction

Arctic climate and hydrology have changed significantly in the past decades. Recently we have studied hydrological regimes and changes over the Kolyma and Lena basins in order to quantify and understand human impact and climatic effects on regional hydrological changes (Yang *et al.*, 2003; Ye *et al.*, 2003; Majhi and Yang, 2008). The Yana basin lies in eastern Siberia and drains into the Mesozoic continental collisional/accretionary zone of very complex geology. This region has a mountainous topography with the Verkhoyansk Range reaching 2000 m. The Yana River is an average sized basin along the coast of the Arctic Ocean. It has a drainage area

[2] *Majhi, I., and Yang, D. (2011): Cold Regions Hydrology in a Changing Climate,* ***39-43.***

of 238,000 km^2 (Fig. 3 . 1), a length of 1,073 km (Huh *et al.*, 1998), and annual discharge of 34 km^3/year. There are no dams in the basin, which has a low population density. It thus provides ideal conditions to examine the effect of climatic variation on streamflow changes. The objective of this study is to better understand factors affecting discharge and its changes. We examine streamflow change and its relation with climatic variables, such as precipitation, temperature and snow cover. The results of this analysis will improve our understanding of hydrological response to climate change in the northern regions.

3.3 Data and Methodology

The Russian Federal Service for Hydrometeorology and Environment (Roshydromet) has monitored discharge of Russian rivers since the early part of this century. The discharge data for this study were obtained from the University of New Hampshire (www.r-arcticnet.sr.unh.edu).

Stage height readings were made daily and cross-channel measurements of discharge were made 25-30 times for a rating curve. Estimates of daily discharge from the rating curves were accurate to ±5% (Shiklomanov and Shiklomanov, 2000). Discharge data are available over various parts of the basins; this analysis focuses on the basin scale. We use monthly and daily flow data collected near the mouth of the river during 1972-1999.

Passive microwave remote sensing in recent years has provided the ability to monitor various features of the Earth's atmosphere and surface, including snowpack properties. A previous study was done on snow covered area for the Lena basin (Yang *et al.*, 2003). A special sensor microwave imager (SSM/I) on the US Defense Meteorological Satellite Program (DMSP) has a daily temporal and good spatial coverage for most areas, which is an important feature for snow pack monitoring. The SSM/I data are from a seven-channel microwave radiometer, which has dual polarized channels at 19, 37 and 85 GHz, and a vertically polarized channel at 22 GHz (Hilburn *et al.*, 2010). The daily SWE data were compiled by the University of New Hampshire from 1987 to 2003, and are available through their website (www.r-arcticnet.sr.unh.edu). We use these data for snow analysis in this study.

The results of the analysis can be briefly summarized in three parts. First, we define the climatic regime and change, using monthly precipitation and data. In the second section, we quantify the discharge regime and change at monthly and annual timescales over the basin. Finally, we compare the snow water equivalent and discharge data at various time scales, in order to examine their compatibility and the effect of SWE change on discharge.

3.4 Result and Discussion

a. Basin climatology

We analyzed temperature and precipitation data for the Yana basin during 1972 to 1999 - a common data period for this study. The mean annual temperature ranged from -14°C to -18°C. The cold temperature is characteristic of regions with continuous permafrost. The basin has a long cold season of eight months, with temperatures ranging from 0°C in September to around -1°C in May. The coldest month is January, with a mean monthly temperature of -45°C. The brief warm season has a temperature range of 9°C in June to 8°C in August; July is the warmest month with a mean temperature of 12°C. Both the seasonal variation and the linear trend have low values. Mean annual precipitation over the basin ranged from 171 mm to 300 mm, with an average of 217 mm. Trend analysis of the precipitation data showed no significant change for any month; the highest change was for July, about 4 mm during 1972-1999. The rest of the months had very low trend.

The month with the most significant trend was April, with an α value of 0.03. It is interesting to note that the precipitation changes are mostly during the winter months, except for August. We compared the relationship between precipitation and temperature during 1972-1999, and found that they were strongly correlated, and significant at an α value of ± 0.05 for a few months. The winter months (October to April) showed positive correlations, implying warmer temperatures associated with higher precipitation.

3.5 Basin hydrology

The Ubileinaya station, situated on the main river valley, (70.77°N, 136.08°E) (Fig. 3 . 1), is the closest station to the basin outlet. It has a drainage area of 224,000 km^2, with a mean annual flow of 1,020 m^3/s from 1972 to 1999. The highest discharge takes place in June (4,300 m^3/s), followed by July (3,055 m^3/s), August (2,615 m^3/s), and September (808 m^3/s). Low flows were from December (12 m^3/s) to April (0.1 m^3/s). Seasonal flow fluctuations showed a high standard deviation for the months with high flows, i.e. June (1,771 m^3/s), July (1,051 m^3/s), August (1,075 m^3/s), and September (805 m^3/s). Flow records showed a decrease of 500 m^3/s in May, and an increase in June by approximately 1,000 m^3/s. The increase in June was compensated by July, which had a decrease of 1,000 m^3/s. Flows in August, September and October also increased by 100 m^3/s (Fig. 3.2). Flow changes in March (-1.5 m^3/s) and April (2.2 m^3/s) were significant. There is a positive change in the base flow for the rest of the winter months. Annual discharge increased by 492 m^3/s (57%) during 1972 to 1999 (Fig. 3.2), significant at 83%; this significant change is most likely due to natural causes.

3.6 SWE *vs* runoff

The snow water equivalent (SWE) data used for this work relate to the period 1988-2000. We examine the relationship between SSM/I snow water equivalent and discharge data collected near the basin outlet. Snow starts to accumulate from September (5 mm) and continues to build up through October (26 mm), November (76 mm), December (116 mm), January (130 mm) and February (Fig. 3.3(b)). The maximum SWE occurs in March, ranging from 133 mm to 180 mm. Max SWE has an increasing trend during 1988 to 1999, with a sudden drop in 2000 (Fig. 3.3(a)). Snow ablation starts in April, when SWE reduced from 146 mm in March to 130 mm in April. With the increase in air temperature in May, SWE reduced to 66 mm. There are variations in snowmelt processes among the years. Earlier melt was observed in the springs of 1997, 1998 and 1999, when the snow had gone by Julian day 142. For the other years, snow disappeared around Julian day 150. In spring 2000, snowmelt was complete by day 150.

3.7 SWE *vs* discharge

Discharge and SWE follow an inverse relationship, with the advent of snowmelt at about day 86 (last week of March) and its final disappearance at day 150, around the last week of May. The discharge subsequently peaks on Julian day 160 (Fig. 3.4). This is typical for the Arctic regions with continuous permafrost. The time series of discharge and snow water equivalent emphasizes the inverse relationship (Fig. 3.5). A high value of SWE does not always lead to a high peak discharge; this could be due to different ablation rates, which are very variable from year to year. We calculated the ablation rates around mid-May to the first week of June, and it varies from as low as 2 mm/d to as high as 16 mm/d at the start of snowmelt. The melt rates change with increase in temperature.

3.8 Conclusion

Yana basin is a pristine and permafrost basin in eastern Siberia. It drains into the Laptev Sea. The basin has a long cold season of eight months, with temperatures ranging from 0°C in September to around -1°C in May. The basin has warmed up over the past three decades. Mean precipitation for the Yana basin ranged from 171 to 300 mm. Precipitation and temperature during 1972-1999 were strongly correlated (significant at α value of ± 0.05) for a few months. The winter months (October to April) showed positive correlations, implying warmer temperatures related with high precipitation. Monthly flow increased during the low flow months and decreased in June. These changes are significant and consistent with the slight increase in temperature over the last three decades. The snow water equivalent shows the maximum accumulation in March; the inter-annual variability in the maximum SWE shows an increasing trend during 1988-2000. The SWE and discharge relationship does not always show consistency, that is, the years with high SWE do not always have peak discharge. Different melt rates could result in this discrepancy. There is a logarithmic relation between daily maximum SWE and discharge. Similar results exist for monthly flow and SWE. It is necessary to continue this research so as to better understand the discharge dynamics and hydrological cycle for the Arctic regions.

3.9 Figures

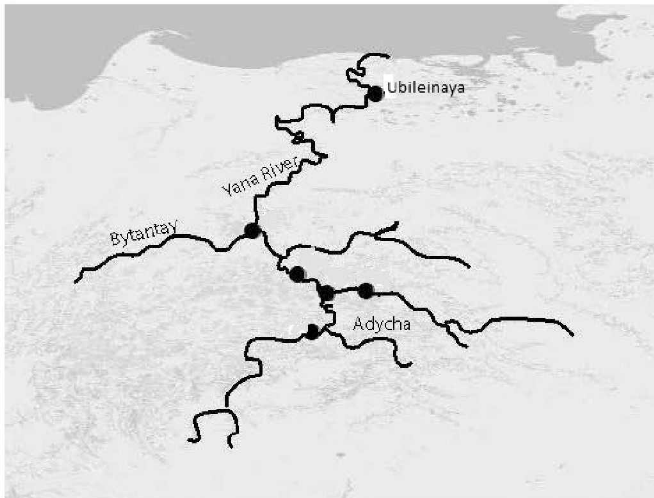

Figure 3.1. Yana basin with all the stations listed from downstream to upstream.

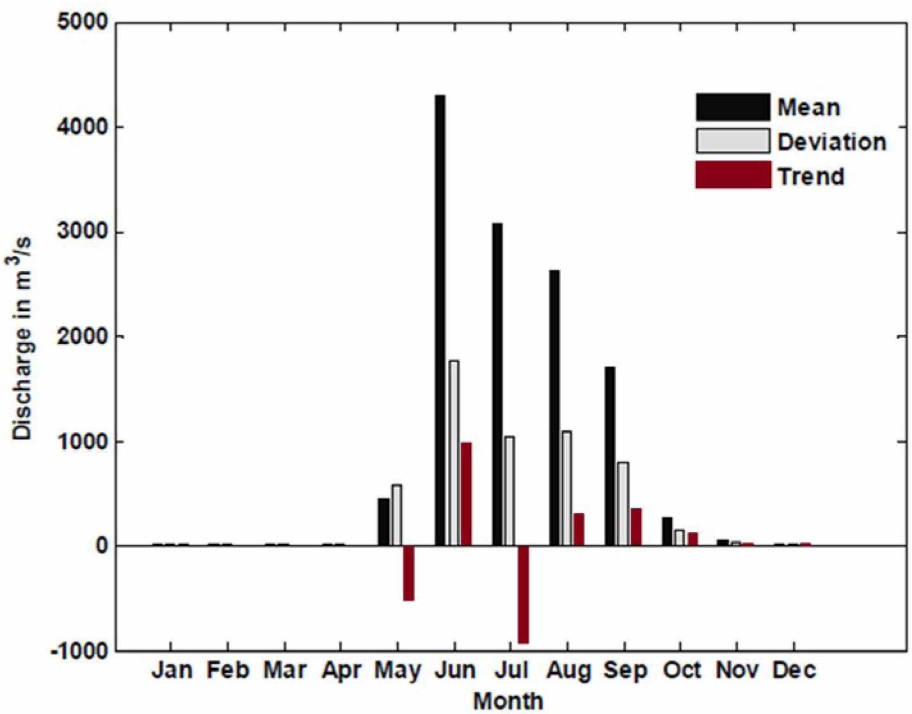

Figure 3.2. Ubileinaya mean discharge, trend and standard deviation from 1972 to 1999.

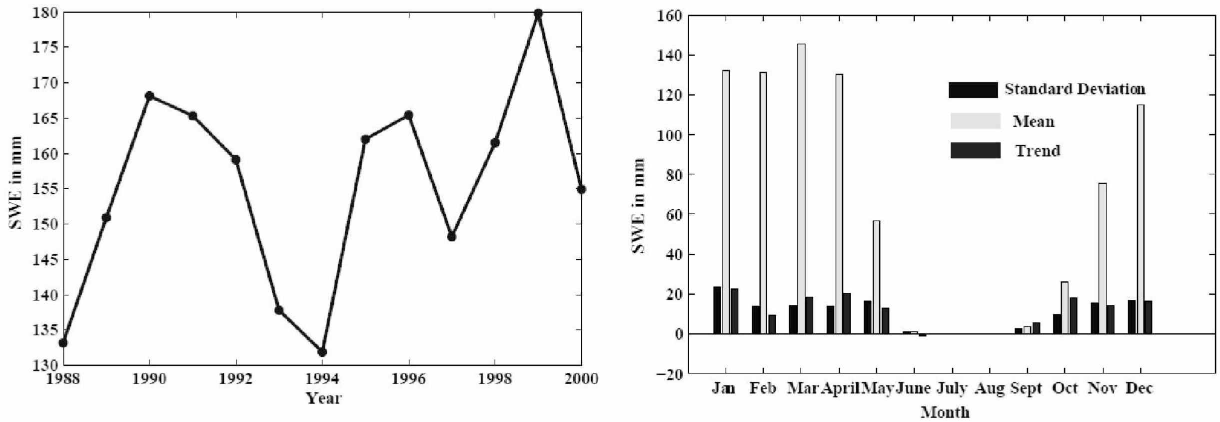

Figure 3.3. (a) Mean March SWE during 1988–2000; (b) monthly mean, STD and trend for SWE for 1988–2000.

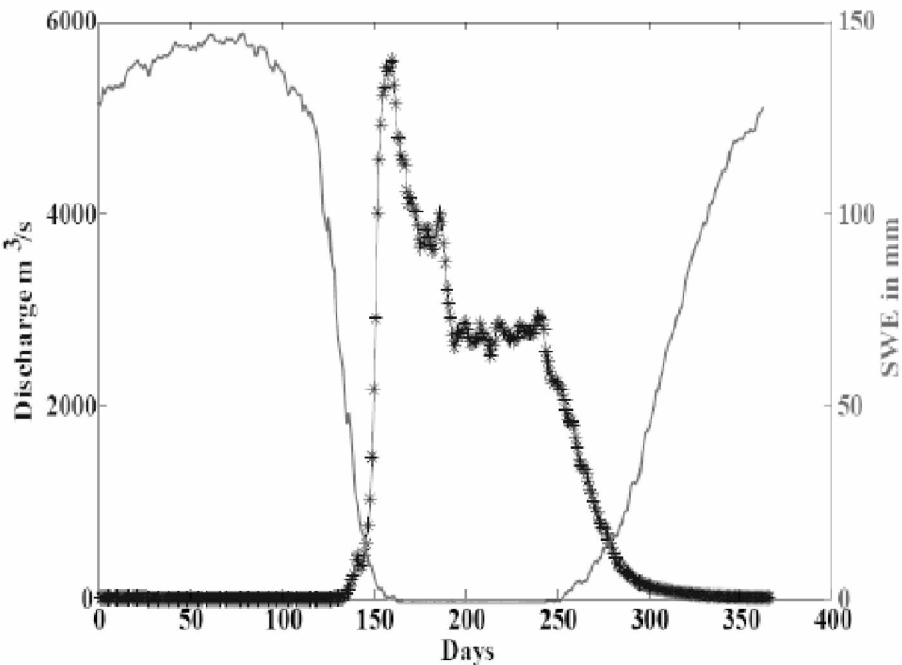

Figure 3.4. Mean daily discharge and SWE for the basin, 1988-2000.

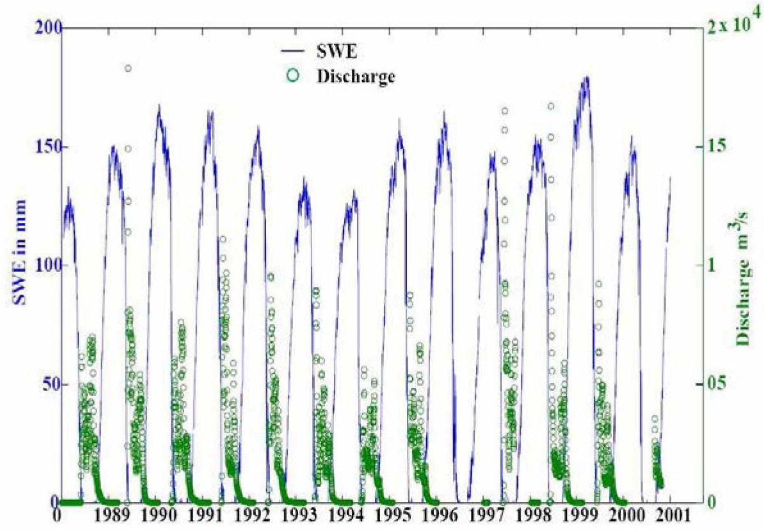

Figure 3.5. Time series of SWE and discharge for the basin, 1988–2000.

3.10 References

Hilburn, K., Wentz, F., Mears, C., Meissner, T. & Smith, D. (2010) Description of Remote Sensing Systems, version 7. *Geophysical Retrievals*, Remote Sensing Systems.

Huh, Y., Ponteleyev, G., Babich, D., Zaitsev, A. & Edmond, J. M. (1998). The fluvial geochemistry of the rivers of Eastern Siberia: II Tributaries of the Lena, Omoloy, Yana, Indigirka, Kolyma and Anadyr draining the collisional/accretionary zone of the Verkhonyonsk and Cherskiy ranges. *Geochimica et Cosmochimica Acta* **62**, 2053–2075.

Majhi, I. & Yang, D. (2008) Streamflow characteristics and changes in Kolyma Basin in Siberia. J. Hydrometro. 9, 267–279.

Shiklomanov, I. A. & Shiklomanov, A. I. (2003) Climatic change and the dynamics of river runoff into the Arctic Ocean. Water Resour. 30, 593-601.

Yang, D., Robinson, D., Zhao, Y., Estilow, T. & Ye, B. (2003) Streamflow response to seasonal snow cover extent changes in Siberian watersheds. J. Geophys. Res.-Atmos. 108(D18):4578, doi:10.1029/2002 JD003149.

Ye, B., Yang, D. & Kane, D. L. (2003) Changes in Lena River streamflow hydrology: human impacts versus natural variations. Water Resour. Res. 39(7), 1200, doi 10.1029/2003 WR 001991.

4. Is there a Link Between Changing Indian Monsoon Seasonality and the Cryosphere?[3]

4.1 Abstract

The Indian subcontinent has recently experienced increases in rainfall during the pre-monsoon months of May and June. The Indian economy is dependent on the monsoon rainfall; therefore, it is important to fully understand the processes associated with variations in this phenomenon. In this study, we explore a potential connection between the cryosphere and changes in rainfall in India. In particular, the main cause of the rainfall increases in May-June has been linked to a general decrease in snow depths in the Himalayas and southern Eurasia as well as an increase in snow cover in northern Eurasia. We track a linkage between an increase of snow water equivalent (SWE) in northern Eurasia and a corresponding decrease in southern Eurasia. Associated with the large-scale SWE anomalies there is a poleward shift in storm tracks and an increase in storm count in northern Eurasia. The reduction of sea ice extent in the fall has been found to be one of s the reason for increased snow in Eurasia and we propose that it changes the storm tracks. In sum, we present evident that is consistent but not conclusive with the notion that large-scale snow cover changes are one of the drivers of increased May-June rainfall in India.

4.2 Introduction

The Indian summer monsoon (ISM) is one of the most energetic and vigorous regional monsoon systems and exhibits highly complex spatio-temporal variability during June-September (JJAS)

[3] Ipshita Majhi[1], Uma S. Bhatt[1], V. Krishnamurthy[2], M.S. Shekhar[2], Soumik Basu[3], Peter Beinek[3] (2018) Is there a Link Between Changing Indian Monsoon Seasonality and the Cryosphere?

(Goswami 2012). It is also one of the largest seasonal anomalies of the global climate system, impacting the lives and economic well-being of millions of people in India through food production and water supply (Shukla 2007). India has an agrarian economy which accounts for 18% of the GDP and the employment of 52% of the workforce in 2014 (Arjun 2013). Thus, any change in climatic drivers of the monsoon season ultimately impacts the farmer along with the economy of Asian countries. Small fluctuations in precipitation can have a large impact on the agriculture and economy of the region by disturbing the overall water cycle (Krishna Kumar et al. 2004; Zhang et al. 2011).

The Indian monsoon is a coupled atmosphere–land–ocean system that is driven by solar heating and is strongly influenced by sea surface temperature (SST) (Krishnamurthy and Krishnamurthy 2014). The land heats up more quickly than the surrounding ocean and the ensuing temperature gradient drives moist maritime winds to converge over the continent, leading to the onset of a monsoon. Understanding the variability in the differential heating between the land and sea helps to determine the strength of the overall monsoon (Bhutiyani et al. 2010; Chatterjee et al. 2009; Das et al. 2002). There are various factors which impact the Indian monsoon such as ENSO, sea surface temperature, and aerosols, to name a few. This study focuses on May- June rainfall, which has notably increased in parts of northern India. This area of India is semi-arid and features an intense low pressure during spring months which leads to dust storms (Narayanan et al. 2013). The month of May is usually characterized by a reversal of surface winds to bring warm, moist and unstable air from Indian Ocean onto the Indian continent (Chandrashekhar 2010).

Aerosols are one of the factors that impact pre-monsoon rainfall in India. One of the theories is that aerosols are thought to reduce clouds, which subsequently increases short wave radiation, increases surface heating, and leads to more June precipitation (Bollasina et al. 2008). Rainfall in the pre-monsoon period is also associated with the formation of semi-permanent heat lows over the northwest parts of India and central parts of Pakistan during the summer months (Chandrashekhar 2010). The recent rise in temperature has led to an intensification of convective heating which causes higher rainfall during March, April and May.

The monsoon is sensitive to global warming (Kitoh et al. 2013) so it is expected that increased GHG emissions will likely increase the severity and complexity of the Indian Summer Monsoon. Climate extremes can catalyze societal instability, such as the civil war like that sparked by a multiyear drought in Syria (Gleick 2014). Over the Indian subcontinent there could be disastrous impacts for agricultural production, ecosystems, health and food security, and the economy (Gadgil and Kumar 2006). This study investigates the mechanisms behind the recent increase in May-June rainfall and this understanding can contribute to improved predictability of the Indian Monsoon. This study will make use of all available data: station observations, remote sensing data, reanalysis products, and global climate model simulations. The available climate model simulations provide insight into the monsoon spanning from the historical to the future period.

Realistic projections of the future behavior of Indian Summer Monsoon (ISM) variability is critical for sustainable economic development and to better adapt in the future. In a future warmer climate from increased GHG concentrations, most studies find an intensification of monsoon rainfall (Meehl and Washington 1993; Hu et al. 2000; Kripalani et al. 2007; Turner and Slingo 2011; Stowasser et al. 2009; Cherchi et al. 2011). Understanding mechanisms behind the monsoon, the different parameters which impact it and how it will change in the future, will help to better predict the monsoon.

This study investigates the possible connection between increases in May-June rainfall with the cryosphere**.** The relationship between the Indian monsoon and Himalayan snow has been known for over a century. Blanford (1884) demonstrated a link between the amount of Himalayan snow cover and the Indian summer monsoon. His proposed mechanism can be described as follows. Above normal snow cover delays heating of the Indian subcontinent and leads to weaker than average monsoons. Conversely, when the snow cover is below average, the land-surface warms up faster and the monsoon is stronger than average. Bamzai and Shukla (1999) found that western Eurasia winter snow cover has a significant inverse correlation with Indian summer monsoon rainfall. This mechanism is similar to the mechanism described by Blanford—that is, snow reduces the land sea contrast. Numerous studies have investigated the inverse relationship between Eurasian snow and the Indian Monsoon (Hahn and Shukla 1976; Dey and Bhanu Kumar 1982; Dickson 1984; Bamzai and Shukla 1999; Kripalani et al. 2007). Several modeling studies have also investigated this snow-monsoon mechanism (Bamzai and Marx 2000; Turner et al. 2007; Saha et al. 2013).

We go a step further to examine if the changing Himalayan snow pattern is linked to larger scale Eurasian cryospheric variability in snow and Arctic sea ice. There is a growing body of research that supports the notion that Arctic sea ice decline drives mid-latitudes extreme weather. Extreme winters in Northern Hemisphere mid-latitudes have been connected to declining Arctic sea ice and continental snow cover changes in autumn through modified planetary waves in the coupled troposphere-stratosphere system (Handorf et al. 2015). Another study linked the recent warming in the Arctic and decrease in sea ice with anomalously large snowfall in large parts of North America, Europe, and east Asia through variations of the Arctic Oscillation (Liu et al.

2012). Sea ice decline is associated with warmer ocean surface waters and intensified turbulent fluxes of heat and moisture into the atmosphere (Cohen and Jones 2011).

Snow depth has an important hydrological effect of enhancing soil moisture, which persists into the following summer and changes the dynamics of evaporation and river discharge (Park et al. 2013). Cyclones are connected with snow in Eurasia which follow a preferred path or track in the extratropics (Whittaker and Horn 1984). A systematic shift in either the geographical location or the intensity of the storm track will result in precipitation anomalies, which are relevant for the regional climate (Chang et al. 2002). Studies show that declining sea ice in the Barents and Kara Seas (BKS) acts as a moisture source for the enhanced Western Siberian snow depth as a result of altered tropospheric moisture transport. There are atmospheric disturbances that are generated due to low sea ice in Barents Kara sea which enter Eurasia from the BKS region and then move southward along the Ural Mountains merging into the extension of the Mediterranean storm track (Wegmann et al. 2015) bringing more snow to the northern part of Eurasia.

In this study, we want to understand a possible pathway by which sea ice variations impact the Eurasian cryosphere and subsequently the Indian Monsoon (Figure 4.1).

Studies have shown that changes in Arctic sea ice extent have wide-ranging impacts, including an increase in snow cover in Eurasia. For example, some parts of Siberia even show an increase in snow depth (Ghatak et al. 2012), while the western Himalayas show a decrease in snow depth (Shekhar et al. 2010). We start our research by asking the few basic questions:

The loss of Arctic sea ice has caused wide-ranging impacts, some of which appear to be related to changes in snow depth and snow cover over Eurasia. In this study, we seek to better understand changes in monsoon seasonality. First, there has been a change in the seasonality of the Indian monsoon (Figures 4.2 and 4.3): positive trends during May-June and negative trends during July-August. Therefore, we focus on the following three questions.

Research Question 1: Are changes in Himalayan snow cover consistent with the recent shift in the seasonality of the monsoon?

Research Question 2: Is there a link between Himalayan snow and Eurasian snow? Can storm tracks help to uncover the connection between the snow and sea ice?

Research Question 3: Is there a link between changes in Eurasian-Himalayan snow cover/depth and Arctic sea ice decline?

To understand the above questions, we will use wide range of data from sources such as station observations, satellite data, model output and CEMS LENS data.

4.3 Data and Methods

4.3.1 Data

4.3.1.1 All India Rainfall Data

The All India rainfall (AIR) index is a long and well-maintained record of monsoon rainfall in India. In this study, we use the area averaged index as well as gridded data that describes rainfall spatially. This study uses the daily gridded rainfall data which has been processed at a high spatial resolution (0.25° × 0.25°) over a period of 113 years (1901-2013) over the Indian mainland (Pai et al. 2014). The gridded data has been qualitatively prepared by the India Meteorological Department (IMD) using the daily rainfall records from the 6,995 rain gauge stations in India. In addition to the spatial data, we have also used the All India time series which is also from 1871- 2014**.** Parthasarathy et al. (1992) created monthly, seasonal and annual rainfall time series for all- India, 29 meteorological subdivisions and 5 homogeneous regions on the basis of fixed 306 rain gauge stations for the period 1871-1990.

4.3.1.2 Snow Water Equivalent

The data for snow water equivalent (SWE) data was obtained from the European Space Agency (ESA) Global Snow Monitoring for Climate Research (GlobSnow) for the northern hemisphere from 1979 to present day (Luojus et al. 2011). The SWE is defined as the amount of liquid water in the snow pack that would be formed if the snowpack was completely melted. The data is a combination of passive microwave radiometer and ground-based weather station data. SWE information consists of terrestrial non-mountainous regions of the northern hemisphere, excluding glaciers and ice sheets. The SWE maps are generated using EASE-Grid projection and a 25-km spatial resolution.

4.3.1.3 Snow cover Extent

The Northern Hemisphere Snow Cover Extent (SCE) was obtained by Rutgers Snow Lab (Robinson 1984) and is based on weekly snow cover maps produced by the U.S. National Oceanic and Atmospheric Administration (NOAA) from visible satellite imagery, including the Advanced High Resolution Radiometer (AVHRR) series and other geostationary and polar orbiting platforms, and in situ data. Their record extends from 1966 to the present and is a global data with resolution of 24 km.

4.3.1.4 Himalaya Snow Depth Data

The Snow and Avalanche Study Establishment (SASE) is a part of Defense Research and Development Organization (DRDO) and is located in Chandigarh, India. SASE maintains weather stations that provide meteorological data and they also conduct weather and avalanche forecasting for the western Himalaya. The observational network of SASE includes 60 surface observatories and three upper-air stations spread over the western Himalaya. The snow depth is measured using a snow stick (which is a plane surface with a one meter measuring stick perpendicular to the surface) every three hours during a snowstorm, otherwise at 0830 hrs and 1730 hrs IST (Shekhar et al. 2010). Figure 4.4(a) shows a map of the study region showing the meteorological stations.

4.3.1.5 Sea Ice

Sea ice data used for this study was provided by the Hadley Center (Rayner 2003). The data spans from 1871 to present and is available at a monthly time resolution. HadISST is the longest gridded sea ice data available for the Arctic. It is a 1° x 1° spatially filled data set which is basically designed as an input for models. HadISST is combination of data from historical ice charts from shipping, expeditions and other observational data, passive microwave satellite retrievals (primarily the NASA Goddard NASA Team data set), and NCEP operational ice analyses (also based on NASA Team and NASA Team 2 algorithms).

4.3.1.6 Storm Tracks

Six-hourly SLP from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) was used to identify storm tracks from 1948-2014. Cyclone, cyclone number and cyclone intensity were calculated using the methodology developed by Zhang et al. (2004). The reanalysis has a horizontal resolution of 2.58° latitude by 2.58° longitude. The cyclone track is defined as the trajectory of the cyclone center and is calculated based on the position of the cyclone and if it moves within radius of 600 km within the 6-h time frame, then it is the new location of the same cyclone and this way all trajectories are calculated, otherwise, a new cyclone is generated. Cyclone intensity was calculated in two steps: first, the mean absolute values of the difference between the central SLP of cyclones and the climatological monthly mean SLP at corresponding grid points over the cyclone duration were calculated for each individual cyclone; second, the mean absolute values were averaged over all cyclones.

4.3.1.7 CESM LENS

The National Center for Atmospheric Research (NCAR) Community Earth System Model Large Ensemble Project (CESM LENS) is a publicly available set of climate model simulations intended to understand internal climate variability and climate change and is described in Kay et al. (2015). The LENS consists of a 40-member ensemble of fully-coupled CESM1 simulations for the period 1920-2100. Each member is subject to the same radiative forcing scenario (historical up to 2005 and RCP8.5), but is initialized with slightly different atmospheric states (created by randomly perturbing temperatures at the level of round-off error). This data set provides many realizations that are very useful to investigate climate variability, in particular natural variations.

4.3.2 Analysis Methods

Standard statistical methods are applied in this study. The simple least squares fit method was used to determine the trends of Indian monsoon, SWE, sea ice and storm tracks. Correlation coefficients were calculated for linearly detrended data to determine the degree to which two parameters vary together. Standard correlation coefficients are used to investigate the relationship between monsoon and Himalayan snow cover, Eurasian SWE and storm tracks. The statistical significance of correlations and trends was assessed using the two-tailed Student's t-test at the 90% or greater level. Composite analysis was conducted for extreme anomalies greater than one standard deviation and used to examine climate anomalies corresponding to anomalously high and low Arctic sea ice states and snow cover.

4.4 Results

4.4.1 Monsoon Trend and Changing Seasonality

This study attempts to sketch a possible pathway between sea ice variations, Eurasian snow cover, and the Indian Monsoon. The change in seasonality of the Indian monsoon with May-June increases and July-August decreases motivates this study (Figure 4.2). The observed increase in May-June rainfall is concentrated in the northern part of India (Figure 4.3). Even though other parameters such as sea surface temperature will always be key, to help better predict the monsoon we look at the cryosphere to provide additional variability. Precipitation trends during May and June over the period 1980-2014 (Figure 4.3) based on the IMD gridded precipitation shows an increase in May and June rainfall especially in the northern region of India. The precipitation increased over northern India (20-30° lat, 70-90° lon) during the 35-year period by 42.55 mm in May and by 10 mm in June. Himalayan snow depth is investigated next to examine whether the Blanford mechanism is operating. Snow depth for Dundhi and Gulmarg (Figure 4.4 c), India, were analyzed for the period of 1990-2015. Dundhi is located about 20 km from Manali at 2,800 m above sea level pressure (Upadhyay et al. 2010) and the observatory at Dundhi (lat. 32.2ºN, lon. 77.07ºE), is surrounded on three sides by high mountains partially covered with vegetation. Gulmarg is a hill station located in Baramula district in Jammu & Kashmir state of India (Figure 4.4a) at 34.06°N and 74.38°E located in north-western Himalayan ranges at around 2,690 m above sea level. It is surrounded by Afghanistan and Pakistan to the north-west and Tibet to the north- east. Gulmarg is located 52 km to the south-west from Srinagar. Snow falls during the months of November to May. There are no major industrial units around this hill station making the site relatively pristine, which is ideal for our study. The major sources of air pollution at this site are tourist activities and emissions from generators used for electricity generation by various hotels (Kumar et al. 2016).

Himalayan snow depth for Gulmarg from 1990-2014 displays a negative trend for both the month of March and April (Figure 4.4c). Dundhi exhibits a consistent trend of decreased snow depths in spring (not shown). These findings are consistent with previous studies that documented snow depth declines from November to April in this region by Shekhar et al. (2010). Trends are negative in February for both stations in this study from 1990-2014 (not shown). Overall, the spring snow depth for both stations has decreased.

As a point of caution, decreases in snow depth could result from warmer spring temperatures and earlier snow melt. The long-term mean temperature trends in the northwestern Himalaya during the 20th century (Bhutiyani et al. 2007) suggest a significant rise in air temperature, with winter warming occurring at a faster rate. Black carbon as a result of pollution also results in reduced snow depths (Ramanathan and Carmichael 2008; Qiu 2008).

Late spring (March-April) snow depth in the Himalayas is negatively correlated with precipitation over northern India is May (Figure 4.5a) and June (Figure 4.5b) as well as over the ocean in May (Figure 4.5c). May and June rainfall is negatively correlated with March snow depth over northern India. The negative correlation is stronger between June rainfall and March snow depth. In addition, March snow depth is negatively correlated with rainfall over parts of southwestern India and over the ocean (Figure 4.5). The correlation is significant in regions with dots at 90%. The decrease in spring Himalayan snow depth is consistent with increased May-June rainfall over India based on the Blandford mechanism. Detrended correlations indicate that the strongest negative correlations are between Himalayan snow depth and north India precipitation in May and June.

4.4.2 Eurasian Snow

The connection between Eurasian snow and Indian monsoon is well documented. In the previous section the Himalayan snow and rainfall in May and June was shown. The next step is to explore if the Eurasian snow co-varies with Indian rainfall in May-June. Snow water equivalent (SWE) in March was correlated with the time series of India May-June rainfall and displays negative correlations in southern Eurasia and positive correlations in northern Eurasia (Figure 4.6). Statistical significance is shown with stippling for levels at the 90% or greater. When spring time rainfall in India is above normal then there is more snow in northern and less in southern Eurasia.

Composite analysis was performed for high and low May-June rainfall from 1980-2014 (See Figure 4.7). Years with anomalies exceeding one standard deviation comprised the high and low cases of the composite analysis. Composites for SWE during October and March are presented in Figure 4.7. Both months show a pattern similar to Figure 4.6, with more snow in the north and less snow in the south in preceding October in years with followed by above normal rainfall for May-June (Figure 4.7).

Variations in storm tracks are explored in the context of snow anomalies. Storm track measures are fairly noisy so are best viewed aggregated over an area. Two sub-regions are delineated to investigate the storm track measures: one over northern and the over southern Eurasia. North Eurasia is defined as the area between 60°-80° N, and longitude 120°-160° W, and South Eurasia is defined over 30°-60° N and longitude 120°-160° W.

We analyzed storm tracks that bring snow to the northern and southern region of Eurasia. Storm track count was averaged over the northern and southern Eurasia domains and were then correlated with the SWE. The results indicate that enhanced SWE in northern Eurasia during October-December coincides with higher storm counts in the north. For the southern Eurasia storm tracks, Figure 4.8c-d the southern Eurasia gets SWE from the southern storm track. In fall, the southern storm tracks move and deposit snow in northern Eurasia. The count of storm tracks in the northern Eurasia matches the SWE for most of the years, with both the SWE and storm track count displaying positive trends (Figure 4.8a). From the above analysis, the results show that snow and storm track is connected. During fall, the north and south storm tracks deposit snow in their respective region. As winter progress and spring arrives, the southern storm tracks along with northern tracks deposit snow in northern Eurasia.

After establishing the relationship between storm track and snow, we further analyze the storm track and intensity. The storm track intensity shows increase from 1948-2014, it does not show much variability between 1980-2014, so we examined the storm count and the storm count for northern Eurasia shows an increase for the fall months of October, November and December and is consistent with the increase in SWE (Figure 4.9b). To understand the decrease in snow in Himalaya, snow cover extent along with snow depth was investigated. Snow cover extent shows a decrease in the fall months of October, November and December (OND) months. The western disturbances which are storms that flow through India from November to May are analyzed. The western disturbance count is decreasing from November-April (Figure 4.9d).

The composite analysis is done for the years with high and low Kara sea ice from 1980-2014. The result is then used to create composite of (Low-High) years of snow water equivalent and for October and November, the years with low sea ice shows higher snow in central Siberia, for month of November there is a dip from extending from Scandinavia to southern Eurasia, which could explain the increase in storm track number (Figure 4.10).

4.4.3 Future of Monsoon: Comparison with Model and Future Simulations

To understand the mechanism of increased May-June rainfall, model simulations were investigated. CESM LENS modeling output was used to investigate the proposed mechanism. There are 40 ensemble runs for historical simulations that span the period from 1920-2005. The mean monsoon seasonality from the pre-industrial run, the historical run and observations compare favorably. The range of May-June rainfall for the 40 ensembles also compares well to observations. The ensembles are fairly close in value with each other and close to observation (Figure 4.11c). The analysis done with the observed data and satellite data match the model results for average of 40 ensembles. The next step would be to analyze the future run to see how the monsoon will change in the future but is beyond the scope of the current study.

4.5 Discussion

The Indian monsoon and snow relationship has been well documented. The pre-monsoon month of May and June are analyzed in our study. Rainfall in the pre-monsoon period is due to the formation of semi-permanent heat lows over the northwest and northern parts of India and central parts of Pakistan during the summer months (Kumar et al. 2016). The rise in the surface air temperature since 1971 (Krishna Kumar et al. 2004) during the months of March, April and May probably suggests the intensification of the convective heating over different parts of India, along with a positive trend of hot days and nights. The month of May is usually characterized by a major reversal of surface winds that brings warm, moist and unstable air from the Indian Ocean into the Indian Peninsular region under the influence of land sea thermal contrast, which plays a major role in the establishment of planetary scale monsoon circulation over the entire subcontinent for coming months (Chandrashekhar 2010). Results show that Himalayan snow depth has negative trend in Feb, March, April and May (Figure 4.4(c)). Reduced snow depth is consistent with rising temperatures. Long-term trends in mean temperatures over the northwestern Himalaya during the 20th century (Bhutiyani et al. 2007) show a significant rise in air temperature in the northwestern Himalaya, with winter warming occurring at a faster rate. Black carbon also helps to reduce snow depth (Ramanathan and Carmichael 2008; Qiu 2008). In this study, it is shown that the decline in snow is linked to changes in storm tracks and their impact on the western disturbance. In an observational study, Das et al. (2002) found that western disturbances in the Himalaya activate monsoons in certain areas of northwestern India. They also studied trends in the annual pre- monsoon (March–May) frequency of western disturbances and the onset date of monsoon over north India for the period 1971–2000 and found that frequency of May western disturbances has significantly decreased over recent years. Our study finds a similar decrease in Western Disturbances from 1980-2014 from November to April, which can also be a process by which Himalaya snow cover decreases, which can lead to earlier heating of the land and higher May-June rainfall. Now, the question remains whether the increase in rainfall is just connected to the local parameters or is there a connection with the cryosphere?

To answer this question, we looked at the Eurasian snow cover in March and correlated it with the May-June rainfall. The analysis showed that May-June rainfall is positively correlated with northern Eurasian snow and negatively correlated with southern Eurasian snow. Less snow in southern Eurasia increases the land sea temperature gradient over Indian continent and favors a stronger monsoon.

Consistent with this study is one by Bulygina et al. (2009), which found that there was increase in SWE in northern Russia and decrease in southern fringes of Russia. This leads us to conclude that there is more snow falling in northern Eurasia and less in Southern Eurasia and this is heating the land up in southern Eurasia and causing the northern part of India to heat up faster leading to earlier rainfall. The impact of Eurasian snow on the Indian monsoon is well researched. Shukla and Mooley (1987) hypothesized that an excessive snowfall during the previous winter and spring seasons can delay the buildup of the monsoonal temperature gradient, since part of the energy will be reflected and part of it will be utilized in melting the snow or evaporating the soil moisture. This can explain the impact of southern Eurasia snow on May-June rainfall. Vernekar et al. (1995) associated higher snow cover with reduced mid-tropospheric temperature gradient over the Indian peninsula, which leads to weaker monsoon. The shift of snow to higher latitude is causing higher May June rainfall and the probable cause is an increase in storm tracks. A systematic shift in either their geographical location or the level of storm activity will lead to substantial precipitation anomalies which has an impact on regional climate (Chang et al. 2002). With their strong connection to weather, storm tracks play a prominent role in mid-latitude climate dynamics. The cyclone activity of the southern Eurasia is significantly correlated with snow in the Northern Eurasia in the winter months. We looked at the storm count and we find it is increasing during winter months in northern Eurasia. McCabe et al. (2001) found that cyclone centers have increased in the Arctic but have decreased in mid-latitudes, while cyclone intensity increased in both high and mid-latitudes. The storm count and snow water equivalent are almost parallel to each other.

To understand the possible cause for a shift in storm tracks and the role sea ice plays in it, composite analysis was conducted. The composite analysis shows that for years with low Kara sea ice, there is above average snow in the central Siberia especially in the month of November and statistically significant area spans from Northern Siberia and down towards the southern edge of Siberia. Studies have shown that declining sea-ice in the Barents and Kara Seas (BKS) acts as moisture source for the enhanced Western Siberian snow depth as a result of altered tropospheric moisture transport. Transient disturbances enter the continent from the BKS region related to anomalies in the planetary wave pattern and move southward along the Ural Mountains where they merge into the extension of the Mediterranean storm track (Wegmann et al. 2015).

This study provides a body of evidence to suggest that sea ice could lead to an increase in May June rainfall by impacting snow. Climate models are useful tools to simulate present climate and provide quantitative estimates of the future climate variability. Over the last two decades, several modeling studies have addressed the link between sea ice decline and Eurasian snow using different sensitivity experiments. Most of these modeling studies advocate the role of increased preferable thermodynamic conditions over ISM region for the intensification of monsoon rainfall leading to the 'wet-get-wetter' situation (Held and Soden, 2006) in response to greenhouse warming (Bhaskaran and Mitchell., 1998; Kitoh et al., 1997; Hu et al., 2000; Turner et al., 2007; Cherchi et al., 2011). The combination of warmer Indian Ocean along with enhanced low-level moisture convergence play a significant role in increasing monsoon rainfall despite the weakened south- westerly monsoon flow through reduction in meridional thermal gradient (Dairaku and Emori, 2006; Meehl et al., 2007; Stowasser et al., 2009; Turner and Slingo, 2011). In the future, the hydrological cycle will invariably be impacted by decreases in Arctic sea ice, increases in snow in the northern Eurasia and decreases in snow in the Himalaya and southern Eurasia.

4.6 Conclusion

The growing populations and the development of socio-economic criteria in India are predominantly interlaced with timing of rainfall; pre-monsoon and monsoon both are significant. Any change in rainfall will have important implications on water resources, agricultural output, public health and economy of the subcontinent. In this study, we have explored the probably causes for increase in May-June rainfall. More than a century ago Blanford suggested an inverse relationship between Himalayan spring snow accumulation and subsequent summer monsoon rainfall over India. We explore the possibility of shift in seasonality by investigating recent changes in Himalayan snow depth and snow cover extent. The results showed the snow depth in March is negatively correlated with May June rainfall in northern India in both May and June and over the two oceanic branches over the Arabian Sea and Bay of Bengal during May. This result reinforces the important role of Himalaya on monsoon rainfall. In this study we also examine how the cryosphere impacts the increase in May June rainfall through changes in Eurasian snow. The March snow water equivalent has a positive correlation with May-June rainfall northern Eurasia and a negative correlation in the southern part of Eurasia. Below average snow depth in southern Eurasia helps to strengthen the meridional temperature gradient and favors earlier rainfall. More snow in northern Eurasia and less in Himalayas and southern Eurasia was shown to be linked with a northward shift in storm tracks. The intensity of storm tracks is increasing in the northern Eurasia and western disturbances, which bring snow in Himalaya are decreasing. The Storm tracks in turn are impacted by Kara sea ice, where less sea ice causes increased snow amounts by deepening cyclones and providing extra moisture. Though the historical conditions do not give a statistically significant positive trend pre-monsoon or shows a strong shift so we look at the future scenario for answers. The CESM LENS ensemble runs when averaged for 40 ensembles reinforce our present result. For future research, a modeling study would provide the answers to missing gap and would help provide better link between the shift in seasonality of monsoon and the cryosphere.

4.7 Figures

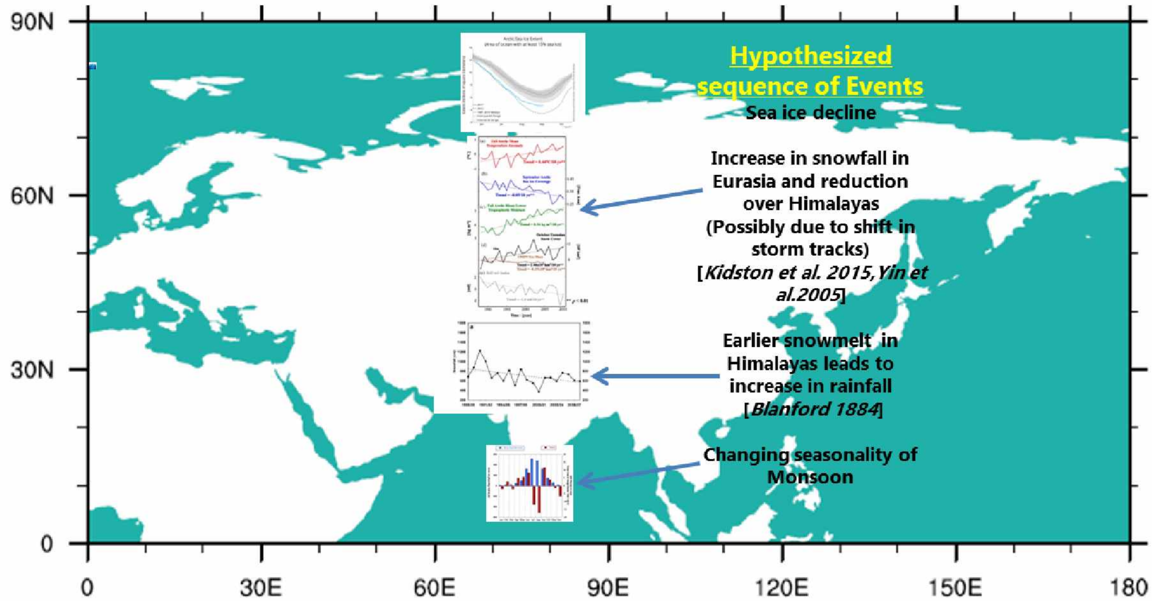

Figure 4.1. Hypothesis: Increase in May June rainfall is connected to cryospheric anomalies in the Himalayas, Eurasia, and Eurasian sea ice.

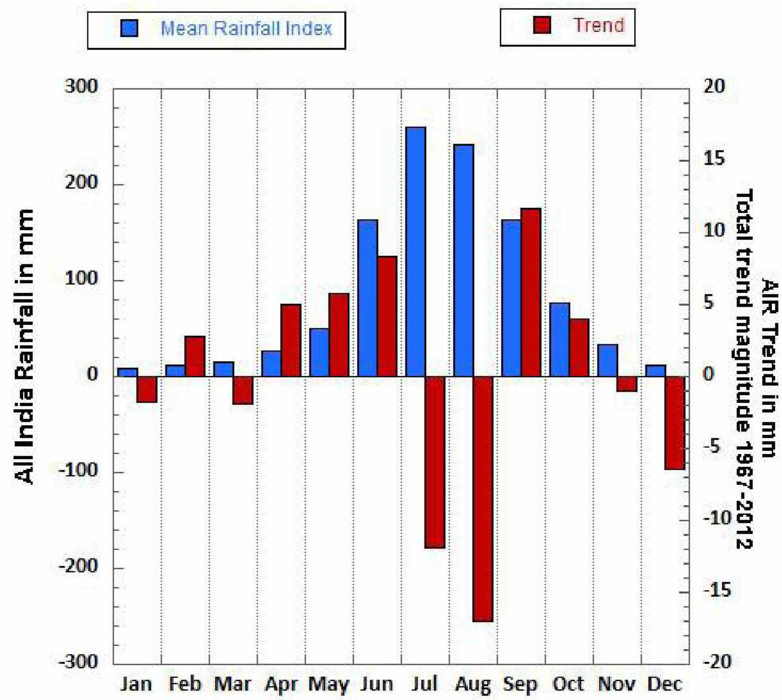

Figure 4.2. All India Rainfall (AIR) monthly mean in mm (blue) and trend (red) in mm over the period of 1967-2014. AIR is in mm and AIR trend is in mm per 46 years.

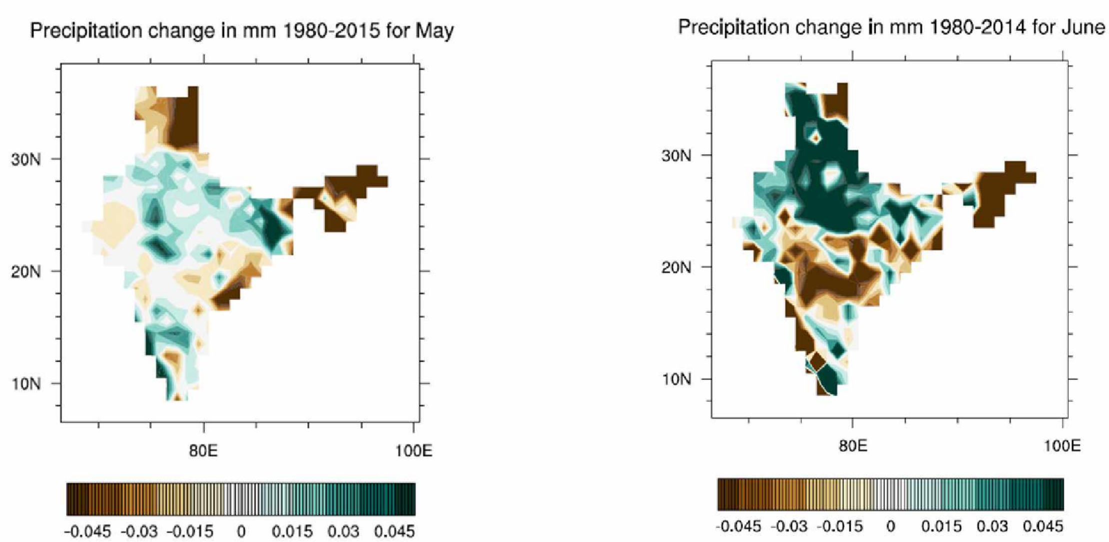

Figure 4.3. Increasing trend for (a) May (b) June rainfall in the northern part of India, using IMD spatial data for India. Precipitation is in units of mm per X and Y years in panel a and b.

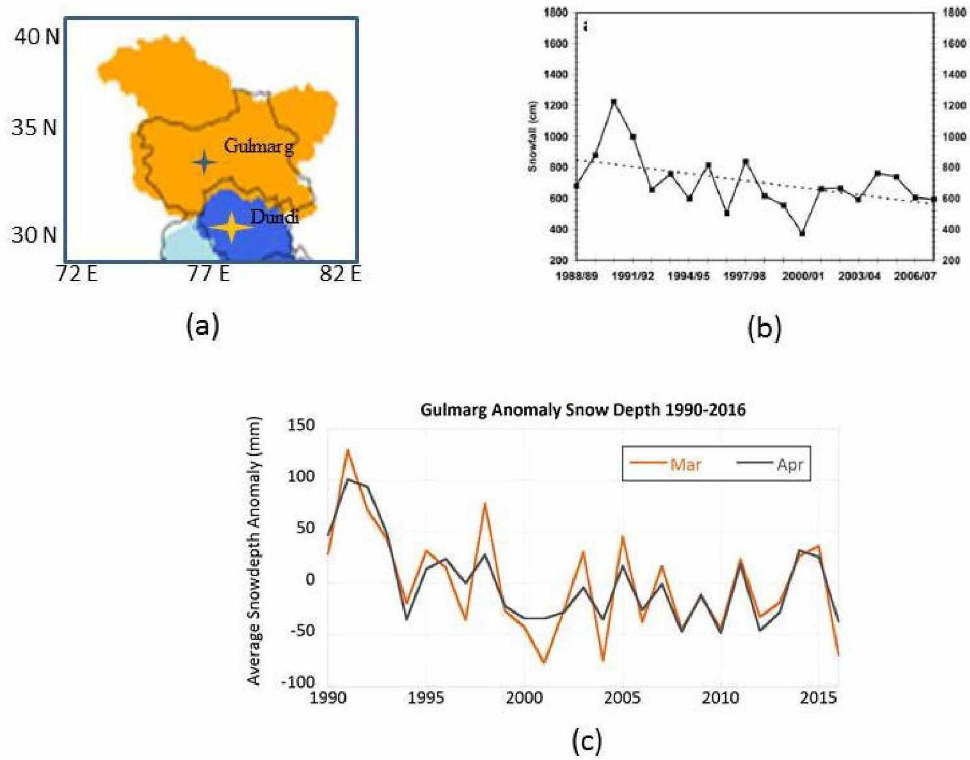

Figure 4.4. (a) Western Himalaya map with location of Gulmarg and Dundhi (b) snow depth from Nov. to April for whole Western Himalaya (panel is Fig 4.a from Shekhar et al. 2010) (c) snow depth trend for Gulmarg for March and April.

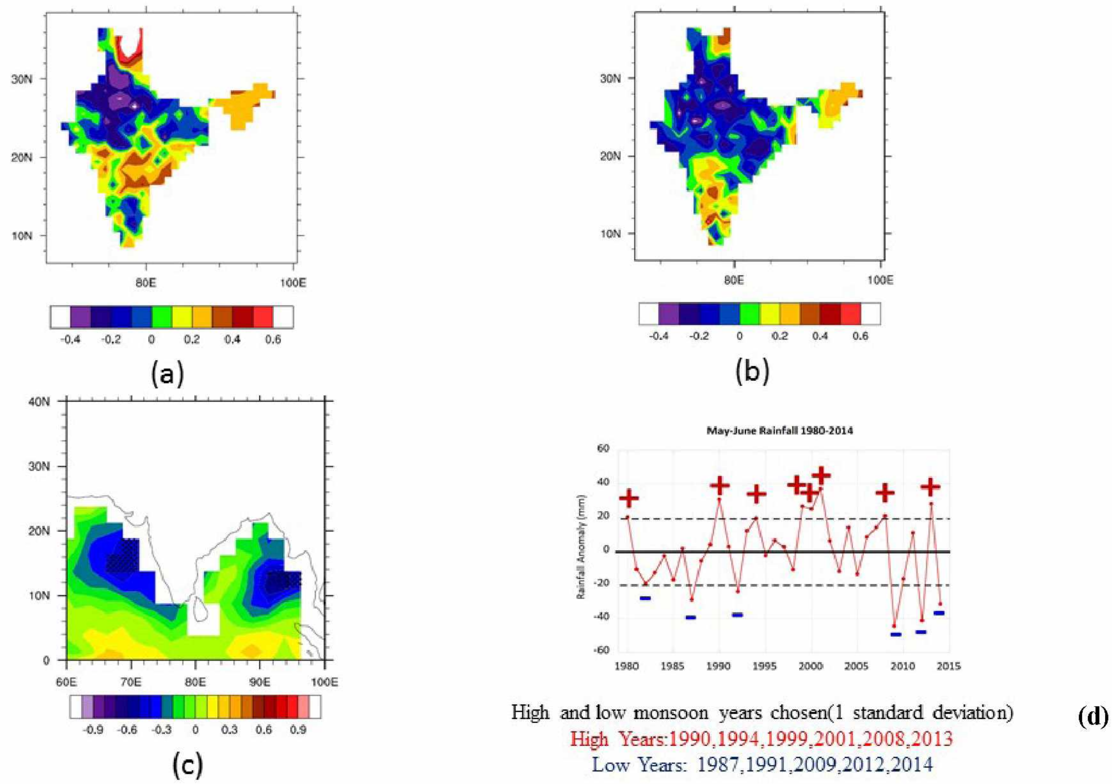

Figure 4.5. Spatial correlation with 90% and higher significance between Correlation between March snow depth in Gulmarg with rainfall in northern India in (a) May (b) June and with rainfall over the ocean in May in (c). (d) Time series from 1980-2014 of May-June rainfall shown in panel (d) with years above one sigma (+) and below one sigma (-) identified for composite analysis.

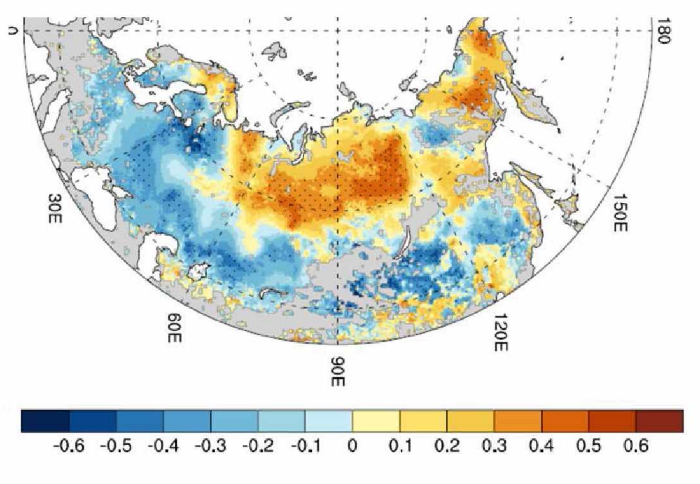

Figure 4.6. Spatial correlation of May-June rainfall with SWE in March. There is a negative correlation with southern Eurasia and a positive with northern Eurasia. Hashing identifies areas of statistical significance at the 90% level of greater.

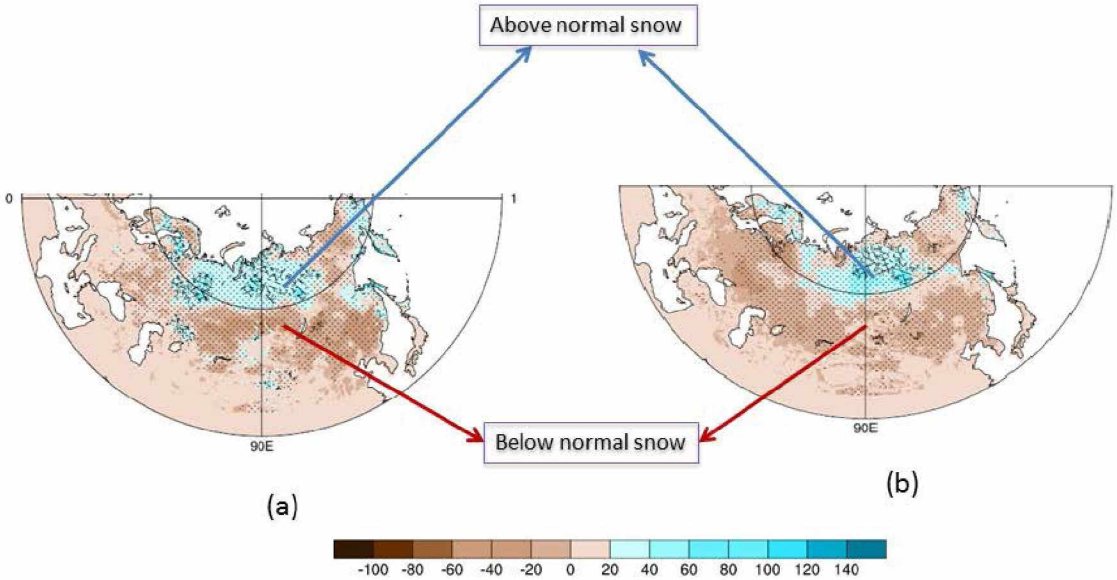

Figure 4.7. Composite (High MJ years-Low MJ years) for SWE in mm for (a) October (b) March. Both panels show higher snow in northern Eurasia and lower in southern Eurasia. Hashing identifies areas of statistical significance at the 90% level of greater.

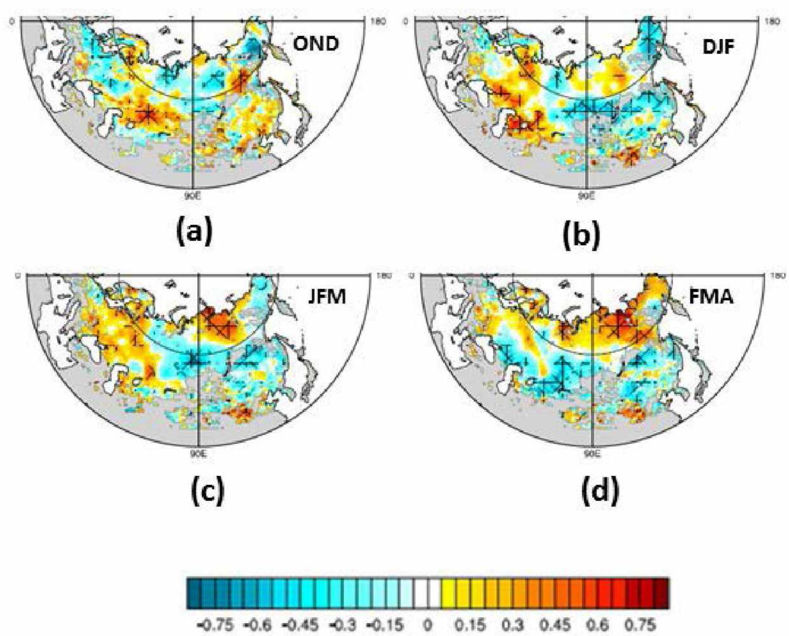

Figure 4.8. Spatial Correlation between northern Eurasia and northern Storm tracks (a) OND (b) JFM. Correlation between southern Eurasia and southern Storm tracks (c) OND (d) JFM. Significance at the 90% level is shown by hashing.

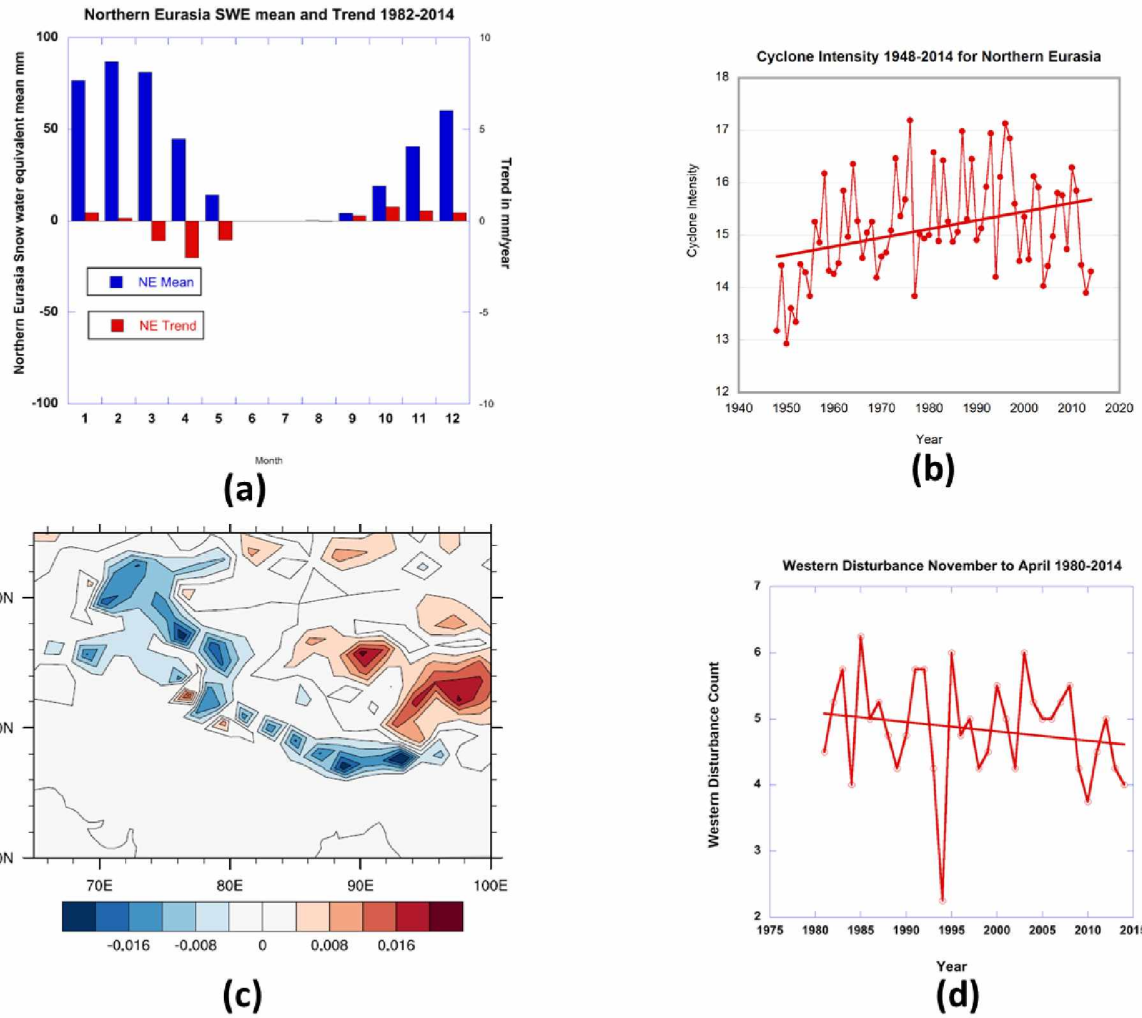

Figure 4.9. (a) Climatology (blue) and trend (red) of SWE for northern Eurasia (b) time series of cyclone intensity in Northern Eurasia (c) now cover extent trend over the Himalaya and (d) time series for western disturbance count for the Himalayas.

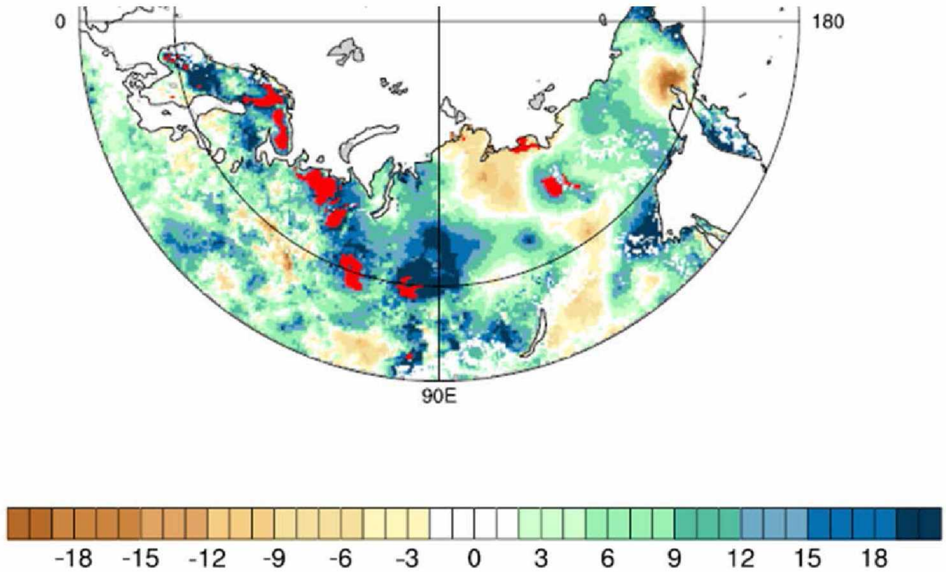

Figure 4.10. Composite of November SWE mm for Eurasia for (Low-High) years for Kara Sea Ice.

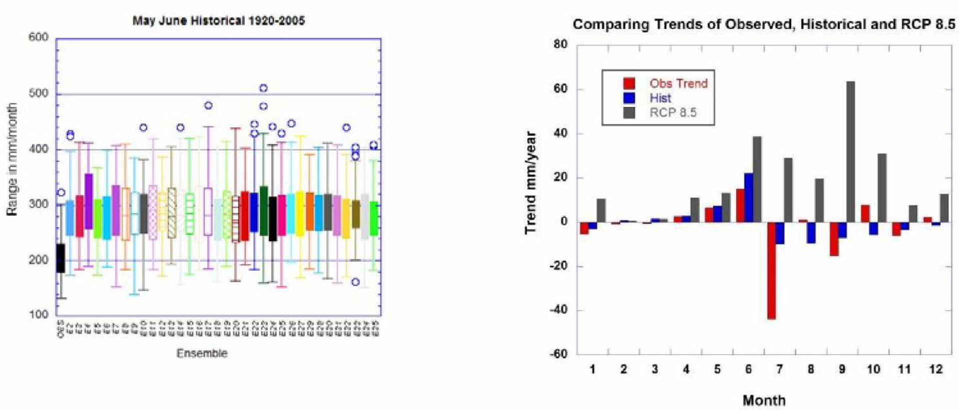

Figure 4.11. (a) The seasonal cycle of monsoon rainfall from the pre-Industrial, historical and RCP 8.5 Ensemble from the NCAR LENS simulations in (a). Select ensembles from the historical simulation are compared to observed (circles) May-June rainfall in panel (b). Trend of May-June rainfall from LENS simulations for the historical (1980-2005) and RCP 8.5 (2005-2080) scenario is shown with observed (1980-2014) trends in units of mm year-1.

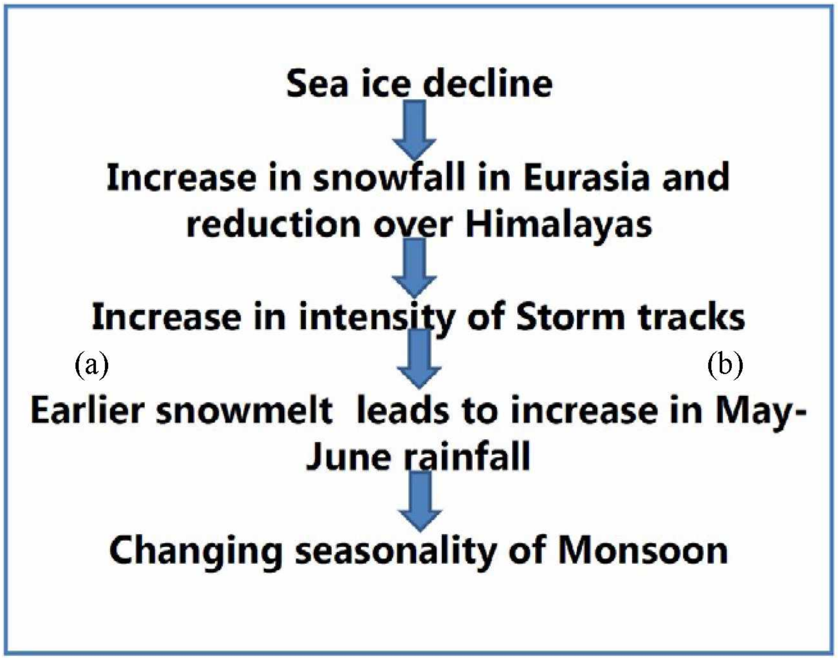

Figure 4.12. Mechanism for connecting the change in Seasonality of Monsoon with sea ice.

4.8 Reference

Arjun, K. M., 2013: Indian agriculture - status, importance and role in Indian Economy. *International Journal of Agriculture and Food Science Technology*, **4**,343–346.

Bamzai, A. S., and J. Shukla, 1999: Relation between Eurasian snow cover, snow depth, and the Indian summer monsoon: An observational study. *J. Clim.*, **12**, 3117–3132.

Bamzai, A. S., and L. Marx, 2000: COLA AGCM simulation of the effect of anomalous spring snow over Eurasia on the Indian summer monsoon. *Q. J. R. Meteorol. Soc.*, **126**, 2575–2584, doi:10.1002/qj.49712656811.

Bhaskaran, B., and J. F. B. Mitchell, 1998: Simulated changes in Southeast Asian monsoon precipitation resulting from anthropogenic emissions. *Int. J. Climatol.*, **18**, 1455–1462

Bhutiyani, M. R., V. S. Kale, and N. J. Pawar, 2007: Long-term trends in maximum, minimum and mean annual air temperatures across the Northwestern Himalaya during the twentieth century. *Clim. Change*, **85**, 159–177, doi:10.1007/s10584-006-9196-1.

Bhutiyani, M. R., V. S. Kale, and N. J. Pawar, 2010: Climate change and the precipitation variations in the northwestern Himalaya: 1866-2006. *Int. J. Climatol.*, **30**, 535–548, doi:10.1002/joc.1920.

Blanford, H. F., 1884: On the Connexion of the Himalaya Snowfall with Dry Winds and Seasons of Drought in India. *Proc. R. Soc. London*, **37**, 3–22, doi:10.1098/rspl.1884.0003.

Bollasina, M., S. Nigam, and K. M. Lau, 2008: Absorbing aerosols and summer monsoon evolution over South Asia: An observational portrayal. *J. Clim.*, **21**, 3221–3239, doi:10.1175/2007JCLI2094.1.

Bulygina, O. N., V. N. Razuvaev, and N. N. Korshunova, 2009: Changes in snow cover over Northern Eurasia in the last few decades. *Environ Res Lett,* **4**:L045026

Chandrashekhar, A., 2010: Basics of Atmopsheric Science. *Basics of Atmospheric Science*, 280–306.

Chang, E. K. M., S. Lee, and K. L. Swanson, 2002: Storm track dynamics. *J. Clim.*, **15**, 2163–2183, doi:10.1175/1520-0442.

Chatterjee, S., S. Ghosh, and U. K. De, 2009: Reduction of number of parameters and forecasting convective developments at Kolkata (22.53°N, 88.33°E), India during pre-monsoon season: An application of multivariate techniques. *Indian J Radio Sp. Phys.*, **38**, 275–282.

Cherchi, A., A. Alessandri, S. Masina, and A. Navarra, 2011: Effects of increased CO2 levels on monsoons. *Clim. Dyn.*, **37**, 83–101, doi:10.1007/s00382-010-0801-7.

Cohen, J., and J. Jones, 2011: A new index for more accurate winter predictions. *Geophys. Res. Lett.*, **38**, doi:10.1029/2011GL049626.

Das, M. R., R. K. Mukhopadhyay, M. M. Dandekar, and S. R. Kshirsagar, 2002: Pre-monsoon western disturbances in relation to monsoon rainfall, its advancement over NW India and their trends. *Curr. Sci.*, **82**, 1320–1321.

Dey, B., O. S. R. U. Bhanu Kumar, 1982: An Apparent Relationship between Eurasian Spring Snow Cover and the Advance Period of the Indian Summer Monsoon, *J. App. Meterol*, **21:12**, 1929-1932.

Dickson, R. R., 1984: Eurasian Snow Cover versus Indian Monsoon Rainfall -- An extension of the Hahn-Shulka Results. *Q. J. R. Meteorol. Soc.*, **23**, 171–173.

Dairaku, K., and S. Emori, 2006: Dynamic and thermodynamic influences on intensified daily rainfall during the Asian summer monsoon under doubled atmospheric CO2 conditions. *Geophys. Res. Lett.*, **33**, doi:10.1029/2005GL024754.

Gadgil, S., and K. R. Kumar, 2006: The Asian monsoon—agriculture and economy. *The Asian Monsoon*.

Ghatak, D., C. Deser, A. Frei, G. Gong, A. Phillips, D. A. Robinson, and J. Stroeve, 2012: Simulated Siberian snow cover response to observed Arctic sea ice loss, 1979-2008. *J Geophys. Res. Atmos.*, **117**, doi:10.1029/2012JD018047.

Goswami, B. N., 2012: South Asian monsoon. *Intraseasonal Variability in the Atmosphere-Ocean Climate System*, 21–72.

Gleick, P. H., 2014: The Syrian conflict and the role of water. *The World's Water*, Vol. 8 of, 147– 151.

Hahn, D. G., and J. Shukla, 1976: An apparent relationship between Eurasian snow cover and Indian monsoon rainfall. *J. Atmos. Sci.*, **33**, 2461–2462.

Handorf, D., R. Jaiser, K. Dethloff, A. Rinke, J. Cohen, 2015: Impacts of Arctic sea ice and continental snow cover changes on atmospheric winter teleconnections. *Geophys. Res. Lett.*, 42, doi:10.1002/2015GL063203.

Held, I. M., and B. J. Soden, 2006: Robust responses of the hydrological cycle to global warming. *J. Clim.*, **19**, 5686–5699, doi:10.1175/JCLI3990.1.

Hu, Z.-Z., M. Latif, E. Roeckner, and L. Bengtsson, 2000: Intensified Asian Summer Monsoon and its variability in a coupled model forced by increasing greenhouse gas concentrations. *Geophys. Res. Lett.*, **27**, 2681–2684, doi:10.1029/2000GL011550.

Kalnay, E. and C., 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Am. Meteorol. Soc.*, **77**, 437–471.

Kay, J. E., and Coauthors, 2015: The community earth system model (CESM) large ensemble project: A community resource for studying climate change in the presence of internal climate variability. *Bull. Am. Meteorol. Soc.*, **96**, 1333–1349, doi:10.1175/BAMS-D-13-00255.1.

Kitoh, A., H. Endo, K. Krishna Kumar, I. F. A. Cavalcanti, P. Goswami, and T. Zhou, 2013: Monsoons in a changing world: A regional perspective in a global context. *J. Geophys. Res. Atmos.*, **118**, 3053–3065, doi:10.1002/jgrd.50258.

Kitoh, A., S. Yukimoto, A. Noda, and T. Motoi, 1997: Simulated changes in the asian summer monsoon at times of increased atmospheric co_2. *Journal of the Meteorological Society of Japan*, **75** (6), 1019–1031

Kripalani, R. H., J. H. Oh, A. Kulkarni, S. S. Sabade, and H. S. Chaudhari, 2007: South Asian summer monsoon precipitation variability: Coupled climate model simulations and projections under IPCC AR4. *Theor. Appl. Climatol.*, **90**, 133–159, doi:10.1007/s00704-006-0282-0.

Krishna Kumar, K., K. Rupa Kumar, R. G. Ashrit, N. R. Deshpande, and J. W. Hansen, 2004: Climate impacts on Indian agriculture. *Int. J Climatol.*, **24**, 1375–1393, doi:10.1002/joc.1081.

Krishnamurthy, L., and V. Krishnamurthy, 2014: Influence of PDO on South Asian summer monsoon and monsoon-ENSO relation. *Clim. Dyn.*, **42**, 2397–2410, doi:10.1007/s00382- 013-1856-z.

Liu, J., J. A. Curry, H. Wang, M. Song, and R. M. Horton, 2012: Impact of declining Arctic sea ice on winter snowfall. *Proc. Natl. Acad. Sci.*, **109**, 4074–4079, doi:10.1073/pnas.1114910109.

Kumar, B., S. Singh, G. P. Gupta, F. A. Lone, 2016: Long Range Transport and Wet Deposition Fluxes of Major Chemical Species in Snow at Gulmarg in North Western Himalayas (India), *Aerosol and Air Quality Research*, **16**, 606–617, doi:10.4209/aaqr.2015.01.0056

Luojus, K., J. Pulliainen, M. Takala, J. Lemmetyinen, C. Derksen, S. Metsämäki, and B. Bojkov, 2011: Investigating hemispherical trends in snow accumulation using GlobSnow snow water equivalent data. *International Geoscience and Remote Sensing Symposium (IGARSS)*, 3772–3774.

McCabe, G. J., M. P. Clark, and M. C. Serreze, 2001: Trends in Northern Hemisphere surface cyclone frequency and intensity. *J. Clim.*, **14**, 2763–2768

Meehl, G. A., and W. M. Washington, 1993: South asian summer monsoon variability in a model with doubled atmospheric carbon dioxide concentration. *Science (80-.).*, **260**, 1101–1104, doi:10.1126/science.260.5111.1101.

Meehl, G. A., C. Covey, T. Delworth, M. Latif, B. McAvaney, J. F. B. Mitchell, R. J. Stouffer, and K. E. Taylor, 2007: The WCRP CMIP3 multimodel dataset: A new era in climatic change research. *Bull. Am. Meteorol. Soc.*, **88**, 1383–1394, doi:10.1175/BAMS-88-9-1383.

Narayanan, P., A. Basistha, S. Sarkar, and S. Kamna, 2013: Trend analysis and ARIMA modelling of pre-monsoon rainfall data for western India. *Comptes Rendus - Geosci.*, **345**, 22–27, doi:10.1016/j.crte.2012.12.001.

Pai, D. S., L. Sridhar, M. R. Badwaik, and M. Rajeevan, 2014: Analysis of the daily rainfall events over India using a new long period (1901-2010) high resolution (0.25° × 0.25°) gridded rainfall data set. *Climate Dynamics*.

Park, H., J. E. Walsh, Y. Kim, T. Nakai, and T. Ohata, 2013: The role of declining Arctic sea ice in recent decreasing terrestrial Arctic snow depths. *Polar Sci.*, **7**, 174–187, doi:10.1016/j.polar.2012.10.002.

Parthasarathy, B., K. R. Kumar, and D. K. Kothawale, 1992: Indian Summer monsoon rainfall indices:1871-1990. *Meteor.Mag.*, **121**, 174–186.

Qiu, J., 2008: The Third Pole. *Nature*, **454**, 393–396, doi:10.1038/454393a.

Ramanathan, V., and G. Carmichael, 2008: Global and regional climate changes due to black carbon. *Nat. Geosci.*, **1**, 221–227, doi:10.1038/ngeo156.

Rayner, N. A., 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, 4407

Saha, S. K., S. Pokhrel, and H. S. Chaudhari, 2013: Influence of Eurasian snow on Indian summer monsoon in NCEP CFSv2 freerun. *Clim. Dyn.*, **41**, 1801–1815, doi:10.1007/s00382-012-1617-4.

Shekhar, M. S., H. Chand, S. Kumar, K. Srinivasan, and A. Ganju, 2010: Climate-change studies in the western Himalaya. *Ann. Glaciol.*, **51**, 105–112, doi:10.3189/172756410791386508.

Shukla, 2007: Monsoon Mysteries. *Science (80).*, **318**, 204–205.

Shukla, J., and D. A. Mooley, 1987: *Empirical prediction of the summer monsoon rainfall over India*.

Stowasser, M., H. Annamalai, and J. Hafner, 2009: Response of the South Asian summer monsoon to global warming: Mean and synoptic systems. *J. Clim.*, **22**, 1014–1036, doi:10.1175/2008JCLI2218.1.

Turner, A. G., and J. M. Slingo, 2011: Using idealized snow forcing to test teleconnections with the Indian summer monsoon in the Hadley Centre GCM. *Clim. Dyn.*, **36**, 1717–1735, doi:10.1007/s00382-010-0805-3.

Turner, A. G., P. M. Inness, and J. M. Slingo, 2007: The effect of doubled CO2 and model basic state biases on the monsoon-ENSO system. II: Changing ENSO regimes. *Q. J. R. Meteorol. Soc.*, **133**, 1159–1173, doi:10.1002/qj.83.

Upadhyay, A., A. Kumar, and A. Chaudhary, 2010: Velocity measurements of wet snow avalanche on the Dhundi snow chute. *Ann. Glaciol.*, **51**, 139–145.

doi:10.3189/172756410791386580.

Vernekar, A. D., J. Zhou, and J. Shukla, 1995: The effect of Eurasian snow cover on the Indian monsoon. *J. Clim.*, **8**, 248–266, doi:10.1175/1520-0442(1995)008<0248:TEOESC>2.0.CO;2.

Wegmann, M., and Coauthors, 2015: Arctic moisture source for Eurasian snow cover variations in autumn. *Environ. Res. Lett.*, **10**, 54015, doi:10.1088/1748-9326/10/5/054015.

Whittaker, L. M., and L. H. Horn., 1984: Northern Hemisphere extratropical cyclone activity for four mid-season months. *J. Clim.*, **4**, 297–310.

Zhang, X., J. E. Walsh, J. Zhang, U. S. Bhatt, and M. Ikeda, 2004: Climatology and interannual variability of Arctic cyclone activity: 1948-2002. *J. Clim.*, **17**, 2300–2317.

Zhang, Q., C. Y. Xu, X. Chen, and Z. Zhang, 2011: Statistical behaviours of precipitation regimes China and their links with atmospheric circulation 1960-2005. *Int. J. Climatol.*, **31**, 1665–1678, doi:10.1002/joc.2193.

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5 Conclusions

5.1 Summary

The Arctic is experiencing unprecedented change from amplified warming and a strengthened hydrological cycle to sea ice decline. Amplified warming is argued by many studies to impact extreme weather in the midlatitudes and possibly beyond to the tropics. It is imperative to better understand different components of the hydrological cycle to improve seasonal to longer-term forecasts to anticipate future changes. This study focused on the Eurasian hydrological cycle from different perspectives: river flow in the Kolyma and Yana basins and the link between Eurasian snow and the seasonality of Indian Monsoon. There exist many gaps in knowledge and this study is aimed to address this problem by studying components of the Eurasian hydrosphere. Climate change and other factors may alter these natural pathways for freshwater, leading to further environmental and societal change in both the Arctic and the tropics.

This study investigated Kolyma River hydrologic regimes and changes induced by reservoir regulation and climate variations. Dams have a major influence on watershed storage, discharge regime, and change. Seasonal discharge changes are stronger and easier to detect than the annual flow changes. This study documents major changes in streamflow hydrology over the Kolyma watershed due to climatic variations and human impacts. Streamflow seasonal cycles over the basin are characteristic of the northern region, with the lowest runoff in April and peak flow in June. Analyses of monthly flows and trends show that reservoir construction and operation have considerably affected streamflow regimes. Comparisons of mean monthly discharge records between pre- and post-1986 indicate that mid-lower basin experience significant increase in low flows and decrease in peak flows after dam construction. In this study we found that low season flow is increasing due to dam constructions, farming, mining, and other activities. This study has improved our understanding of human impacts on discharge which is important to gage the societal impacts of change in discharge and the timing of freshwater input is important since increase in winter flow could stall convection of shelf.

The Yana Basin was chosen to investigate the impact of climatic parameters where there has been minimal human interference with the natural system. The Yana Basin is a nearly pristine basin in Eastern Siberia that drains into the Laptev Sea and is surrounded by continuous permafrost. The basin is characterized by a long cold season of eight months, with temperatures ranging from 0°C in September to around -1°C in May. The basin has warmed up over the past three decades. Mean precipitation for the Yana basin ranged from 171 to 300 mm. Precipitation and temperature during 1972–1999 are significantly correlated. Discharge data near the basin mouth suggest both an increase and decrease over the summer months. Basin precipitation has a positive correlation with discharge during June, July and August. The relationship between snow water equivalent and discharge follows an inverse relation; maximum snow water equivalent and discharge have a linear relation, with inconsistencies in some years. This study leads to a better understanding about the discharge regime that both precipitation and temperature together shape the discharge peak and contributes to better understanding of the hydrological regime and changes over the northern regions.

The final component of the Eurasian hydrological cycle this thesis examined was change in Eurasian SWE and how it could be linked to the Indian monsoon. It is well-established that the strength of the Indian monsoon is affected by the amount of Himalayan snow cover. This thesis documents an increase in rainfall during May and June, which can adversely impact agricultural production and have socio-economic consequences. This study is an attempt to outline a possible mechanism to explain possible drivers of the increase in May June rainfall, namely Himalayan snow, Eurasian snow, and Arctic sea ice. The process is summarized as follows. The decrease in snow amount and earlier snowmelt allows the land surface to heat up over a longer period. This earlier start to the heating of the Indian subcontinent leads to an intensified thermal gradient between the land and the ocean. Consequently, this leads to an influx of moisture from the ocean and promotes an increase in May June rainfall. SWE decline in southern Eurasia in recent years coincides with an increase of SWE in Northern Eurasia. This increase in northern Eurasian snow is coincident with an increase in storm track counts that originate in southern Eurasia, bringing moist subtropical/midlatitude air from southern to northern Eurasia. Composite analysis presented in this thesis suggests that the reorganization of the storm tracks is coincident with sea ice anomalies in the Kara Seas. Years with more open water in September are associated with more snow in northern Eurasia and vice versa. This observational analysis can help to develop a possible mechanism but climate models are needed to show that sea ice is forcing the snow anomalies. To understand the future changes in the monsoon, with the sea ice almost gone, we used CESM LENS model output. The future of monsoon shows a positive trend in all the months accept June, which could be due to a delay in the onset for thermodynamic reasons. Overall these analyses have shown that there many complexities related to change in Indian monsoon or Arctic river discharge. While many gaps in knowledge of change in seasonality of monsoon or discharge still exist, these studies have each helped to bridge a few of these gaps.

5.2 Conclusions

These analyses have shed light on the important changes in the Eurasian hydrological cycle that may link the Arctic and tropics. Evidence presented in this study supports a possible mechanism, though it is not the only possible mechanism. There are other possible mechanisms which were not covered in our study. Kolyma river basin showed how the winter discharge is increasing due to anthropogenic reasons and Yana basin showed how temperature and snow play a crucial role in discharge. Hydrological changes in snow amounts have wide ranging impacts: change in Eurasian snow impacts not only Arctic river discharge but also the Indian monsoon. The Indian monsoon shows an increase in the month of May and June, which was shown to be consistent with variability in Himalaya and Eurasian snow. A physical mechanism was provided that links between climate process in Arctic and the tropics acting on time scales of 1-3 seasons. Any additional information to create better forecasts is important for planning and socio-economic reasons. This knowledge will have better understand climate variability and aid future climate researchers. Based on the results of these studies it can be concluded that winter discharge in some of the Arctic rivers with dams have dramatically increased. The pristine Arctic basin is impacted by change in temperature and solid precipitation. The Eurasian snow along with Himalayan snow can modulate the Indian monsoon.

5.3 Future outlook

The Arctic is undergoing unprecedented change, it is important to model the changes and predict the future outcome for us to better prepare for the future changes whether it be infrastructure or weather. Also, there is a need for better and more fruitful collaboration between scientist in the Arctic, mid-latitudes and the tropics so that the global climate can be viewed as a whole. The moisture transport from lower to higher latitudes needs to be better understood, also changes in the global hydrological cycle has to be understood. Synthesizing modeling studies, remote sensing data, and observational station data is necessary to better understand the impact of changing climate on discharge and Monsoon. Critical gaps in knowledge concerning mechanisms that drive the change in Eurasian and Himalayan snow via change in sea ice and storm track needs to understood. The people who live in the Arctic face great challenges due to its extreme climate conditions and any small change in monsoon impacts billion of people in the tropics. Improving our understanding of how the climate functions helps to reduce some of this uncertainty and improves the human condition.

Finally, a platform for communication between the climate science and users from different communities needs be developed further and created at the international level. In doing so practical applications of research and science will be available for stakeholders that included planners, people and society in general. These steps will help politicians, scientists and different aspects of humanity make better decision in response to current and future changes in weather and climate.