The Global Energy and Water Cycle Experiment (GEWEX) is a program initiated by the World Climate Research Programme (WCRP) to observe, understand and model the hydrological cycle and energy fluxes in the atmosphere, at land surface and in the upper oceans. An important component of this initiative is the Continental Scale Experiments being coordinated by the Global Hydrometeorology Panel (GHP). These experiments were designed to improve our measurement, understanding and ability to model the processes and feedbacks that affect the water and energy cycles of the climate system over different land areas of the earth. This will support the overall GEWEX goal to reproduce and predict, by means of suitable models, the variations of the global hydrological regime, its impact on atmospheric and surface dynamics, and variations in regional hydrological processes and water resources and their response to changes in the environment, such as the increase in greenhouse gases.

The Continental Scale Experiments that are being coordinated by GHP include the GEWEX Americas Prediction Project (GAPP), the Baltic Sea Experiment (BALTEX), the GEWEX Asian Monsoon Experiment (GAME), the Large Scale Biosphere-Atmosphere Experiment in Amazonia (LBA), and the Mackenzie GEWEX Study (MAGS). The associate project, Couplage Atmosphère Tropicale et Cycle Hydrologique (CATCH), also contributes to the realization of the GHP goals.

Figure 1: The GEWEX-GHP continental scale experiments and associate projects.
The MAGS experiment and the Siberian component of the GAME experiment share a concern with the energy and water cycles of cold dry climates (Figure 2). Both are focused on the basins of northward-flowing rivers (the Mackenzie and the Lena, respectively) that lie within regions that are experiencing a significant warming trend. Extensive areas in both basins are underlain by permafrost and share common a vegetation cover (including forests and tundra). Both experiments also face a similar challenge of studying the water and energy cycles with only a sparse network of ground observation stations in a region where access is often difficult and expensive.

As might be expected for two continental scale experiments that share such similarities in landscapes, goals and scientific challenges, collaborative efforts have ensued. One of the most substantial was the 1st GAME-MAGS International Workshop held in November 1999 in Edmonton, Canada. The aims of that workshop were:

1. To establish a better appreciation of the similarities and differences between the two projects.
2. To develop a better appreciation of critical scientific issues facing both of the projects.
3. To consider plans for future joint activities.

The success of that workshop has led to this 2nd GAME-MAGS International Workshop held in October 2001 in Hokkaido, Japan. The overarching goal of this workshop was essentially the same as that of the first workshop; namely, to enhance collaboration between the MAGS and GAME continental scale experiments. This workshop focused particularly on reviewing scientific progress over the past several years in the understanding of the water and energy cycles of cold, dry regions and explored the potential for collaborative research projects. A most encouraging outcome has been the development of a process by which collaborative research project in some targeted topical areas can be initiated. Areas identified for potential collaboration were:
1. Hydrological research
   - Hydrological models
   - Process research in:
     - Tundra environments
     - Boreal forest environments
     - Mountainous environments

2. Atmospheric research

3. Remote sensing

4. Data issues

Another joint workshop is tentatively planned for 2003 to exchange information on the new collaborative research initiatives.

We are grateful to the Institute for Low Temperature Science of Hokkaido University for organizing and providing facilities for our workshop. We also wish to thank Peter di Cenzo of the MAGS Secretariat for his assistance in editing and publishing this proceedings. We are also grateful for the support from Natural Sciences and Engineering Research Council of Canada (NSERC), Environment Canada and Frontier Research System for Global Change.

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ATMOSPHERIC STUDIES
1. Introduction

The Asian winter monsoon (AWM) is one of the most energetic and convectively active systems of the atmosphere. The large thermally driven overturning circulation of the AWM is characterized by two principal components. Over northern Eurasia strong radiative cooling and cold air advection contribute the pronounced Siberian high and its associated heat sink. To the south over the tropical western Pacific, the so-called 'warm pool' region with sea-surface temperatures in excess of 28.5°C, intense convection and heavy precipitation lead to a large condensational heat source. Air flows in a south and southeast direction from the heat sink to the heat source at low-levels, with a return flow in a north and northwest direction at upper-levels from the heat source to the heat sink.

East Asian cold surges are very pronounced high-frequency fluctuations of the AWM, characterized by the rapid intensification and subsequent southeastward movement of the Siberian high, causing cold dry air to surge southward in the vicinity of the South China Sea (5ºN-10ºN, 110ºE-115ºE; see Figure 1). These cold surges move rapidly, spreading equatorward from the extratropics to the subtropics over approximately a 4 day span (Garreaud 2001). Increases in both the area and intensity of convective activity over the South China Sea, the Philippine Sea and the Indo-China coast following pressure surges are well documented (Compo et al. 1999).

Numerous studies have shown that the tropical latitudes over Indonesia and the western Pacific Ocean are preferred regions for interhemispheric interaction, which is enhanced during cold surge events (Davidson et al. 1984; Love 1985; Kiladis et al. 1994).

The onset of the Australian summer monsoon (ASM) has been linked to the occurrence of cold surges from Southeast Asia (Davidson et al. 1984; Suppiah and Wu 1998). On intraseasonal time-scales, the effects of cold surges and the shifting of atmospheric mass equatorward can often lead to an increase in the west to east near equatorial pressure gradients over the tropical western Pacific (Love 1985; Kiladis et al. 1994). The enhanced low-level westerly flow, resulting from the strengthened pressure gradients, acts to increase the cyclonic vorticity within the ASM trough region through the effects of lateral shear (Love 1985; Kiladis et al. 1994). More recent studies have also found a strong connection between these so-called westerly wind bursts in the lower troposphere of the tropical western Pacific and mid-latitudes pressure surges originating over Southeast Asia (Lau et al. 1996; Meehl et al. 1996; Suppiah and Wu 1998).

A study by Murakami and Sumi (1982) examined active and break phases of the southern hemisphere monsoon for the 1978-79 season. The authors found that vorticity advection by the northerly divergent winds were largely responsible for the intraseasonal changes in the zonal mean vorticity and hence the strength of the monsoon. These low-level divergent winds resulted
from fluctuations in the strength of the low-level Hadley circulation which are known to accompany pressure surges.

2. Dry Atmospheric Loss from the Northern Hemisphere

A composite study of 25 significant events of subseasonal dry atmospheric mass loss from the northern hemisphere during the boreal cold season indicated that these events occurred rapidly, possessing a preferred time-scale of 9 days. Pressure surges over Southeast Asia and North America, associated with statistically significant positive atmospheric mass anomalies, were mechanisms which acted to channel the atmospheric mass equatorward on such a rapid time-scale (~4 days).

To summarize the findings from the 25 case composite study, we present in Figure 1 a depiction of the localized regions which undergo atmospheric mass increases and decreases during the significant events of northern hemisphere dry atmospheric mass loss. The atmospheric mass is shown by means of sea-level pressure (SLP) anomalies derived from the recently completed National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) reanalysis. The statistical significance, as calculated based upon a two-sided Student’s T-test, is also shown. Negative (positive) values denote regions that have undergone an atmospheric mass decrease (increase) throughout the composite event.

Virtually the entire area poleward of 45ºN has undergone a loss of atmospheric mass. In particular, three distinct regions in the mid- to high latitudes of the northern hemisphere have experienced an atmospheric mass evacuation: (i) the North Atlantic to the south of Greenland, (ii) the North Pacific, extending southeastward to western North America, and (iii) the northern Asian continent. These regions are favored locations for low-frequency persistent anomalies during the northern hemisphere winter months. In the southern hemisphere the atmospheric mass preferentially accumulates in the South Pacific Ocean to the east of New Zealand and in the South Indian Ocean.

The surging of atmospheric mass southward over Southeast Asia and the decrease in the lower tropospheric temperatures over China combined with the increase in the northerly component of the near surface wind field to the south of China are three prominent characteristics of cold surges over Southeast Asia. A time-series of each of these components is shown in Figure 2 for the 41 day period surrounding the onset time (T0) for the composite northern hemisphere dry atmospheric mass collapse event. The three regions we consider are depicted in Figure 1. The region over northern Eurasia is referred to as the northern box and the region over eastern China is referred to as the southern box in Figs. 2a and 2b. Figure 2a depicts a substantial rise in anomalous atmospheric mass over northern Eurasia prior to the event (solid curve), followed by a more pronounced decrease during the event. The magnitude of the rise prior to the event is of the order of 6 hPa and an analysis of the individual events indicates that the rise is associated with the building of the Siberian high, while the subsequent fall of the order of 8 hPa is related to the southeastward propagation and simultaneous weakening of the cold surge anticyclone.
Figure 1: Composite SLP anomaly difference, contour interval of 1 hPa to a magnitude of 4 hPa, with every 2 hPa for larger magnitudes. For each event the SLP anomaly difference is calculated as the difference between the times of local maximum and the local minimum in anomalous NH dry atmospheric mass for each event. Light (dark) shading indicates the statistical significance at the 95% (99%) level based upon a two-sided student's T-test. The three boxes orientated from north to south refer to the northern box, the southern box, and the South China Sea discussed in the text.

The equatorward surging of the anomalous atmospheric mass is seen in the time-series for the SLP anomalies over eastern China (dash curve; Figure 2a). The anomalous atmospheric mass peaks in this more southerly area approximately 2 days after the peak over northern Eurasia. Figure 2b clearly depicts a large drop in lower tropospheric temperatures over eastern China as represented by the 1000-850 hPa thickness anomaly. The timing of the drop in thickness coincides with the large drop in anomalous SLP over northern Eurasia (Figure 2a). To the south of China in the South China Sea the northerly winds gradually increase in strength after the onset time and peak 6 days after the onset (Figure 2c). Clearly there is a Southeast Asian cold surge signature in the 25 case composite.
Figure 2: East Asian cold surge signatures for the SLP anomaly (top panel) in the northern and southern boxes of Figure 1, 1000-850 hPa thickness anomaly in the northern box (middle), and surface meridional wind anomaly in the South China Sea.
3. Individual Case Study

The analysis of an individual event of dry atmospheric mass loss from the northern hemisphere, which is representative of the 25 case composite, reveals both the direct and indirect roles of a Southeast Asian pressure surge on the collapse of dry atmospheric mass from the northern hemisphere. The event began on 3 March 1989 (onset time) and lasted for 9 days. The total anomalous dry atmospheric mass loss from the northern hemisphere was 1.8 hPa, ranking it 7th among the 25 events. To put this number of 1.8 hPa into perspective, it represents greater than 60% of the mean annual cycle of northern hemisphere dry atmospheric mass.

Associated with this event was a large evacuation of atmospheric mass from northern Eurasia which occurs as the atmospheric mass surges equatorward and into the southern hemisphere. Along the west coast of Australia, a southerly pressure surge extends equatorward and converges with the northerly surge to create a pronounced near equatorial zonal pressure gradient. A low-level westerly wind burst develops within the monsoon trough (10ºS - 15ºS) in response to this enhanced zonal pressure gradient as part of the onset of an active phase of the ASM. Within the ASM trough, the deep layer of westerly winds extends from the surface to 400 hPa and is overlain by easterly winds. A 2 to 3 day time lag is found between the maximum northerly winds over the South China Sea and the onset of the deep westerlies within the monsoon trough of northern Australia. This temporal phasing suggests a strong role played by the pressure surge in initiating or triggering the onset of the active phase of the ASM. In accordance with the onset of the active phase of the ASM deep convection is found over northern Australia beginning 7 March, which spreads southeastwards with time as part of a deepening monsoon trough over central Australia.

Three prominent anticyclonic circulations intensify in the southern hemisphere extratropics, stretching from the South Indian Ocean to the South Pacific. The intensifications occur beneath regions of upper tropospheric dry atmospheric mass convergence, originating from the monsoon convection outflow. The role of the diabatic heating anomalies is to redistribute the dry atmospheric mass in the divergent outflow of the upper troposphere. At the end of the northern hemisphere dry atmospheric mass collapse event, these anticyclonic circulations are largely responsible for the dry atmospheric mass increase in the SH.

4. References


1. Introduction

The atmospheric and surface branches of the hydrologic cycle are coupled through precipitation and evapotranspiration. Precipitation $P$ over a region can be partitioned according to the source of the moisture which contributes to the precipitation: (i) $P_a$, precipitation derived from water vapour advected into the region by atmospheric circulations, and (ii) $P_m$, precipitation derived from water vapour supplied by evapotranspiration within the region. This latter mechanism provides a means for the land-atmosphere moisture recycling or precipitation recycling to occur over the region. A quantitative measure for precipitation recycling over a specific region is given by the precipitation recycling ratio $\rho$ defined as the ratio $P_m/P$ (Eltahir and Bras, 1996). The degree to which moisture recycling is active over a region gives a diagnostic measure of the importance of land surface processes and land-atmospheric interactions in governing the water budgets over the region. In addition, recycling can effectively redistribute the surface water content over long distances within a large region. Depending on the spatial variability of its surface characteristics, this spatial redistribution of the surface water content can have critical effects on both the local and regional-scale hydrologic responses to large-scale atmospheric forcings.

The purpose of this study is to estimate the precipitation recycling ratio as well as its temporal and spatial variability over the Mackenzie River Basin (MRB). Results of the calculations will be interpreted with the meteorological and surface features and processes that affect the region. The response of some of the basin’s important hydrologic components (e.g. precipitation and its recycling, and runoff) to variations in the large-scale atmospheric conditions will be investigated by examining the results for both “normal” and selected “anomalous” years. The MRB moisture recycling study had been recently extended to other major river basins. Some preliminary results obtained for the Lena River Basin will also be presented and compared to the MRB results.

2. Methodology and Datasets

2.1 Estimation of the Precipitation Recycling Ratio

Several empirical formulations have been developed to estimate the precipitation recycling ratio for a specified spatial domain (see Brubaker, 1993, and Eltahir and Bras, 1994 for detailed derivations of some of these formulations). Similar assumptions are typically adopted in deriving these formulae for calculating the recycling ratio: (i) the planetary boundary layer (PBL) is well-mixed; and (ii) the change in atmospheric water vapour storage over long time scales is small compared to the atmospheric fluxes (including the evaporative flux) of water vapour. When these assumptions along with the conservation of water mass are applied to a control volume within the domain, the following equation can be derived (Eltahir and Bras, 1994, 1996):
\[ \rho = \frac{I_i + E}{I_i + E + I_e} \]  

where \( E \) and \( I \) represent the evaporation and water vapour influx for the control volume, respectively. The indices \( i \) and \( e \) denote respectively the variables corresponding to the internal (i.e., from evapotranspiration within the domain) and external (i.e., from advection by airflow into the domain) sources of water vapour. By dividing up the domain of interest into a grid of control volumes, one can apply Eq. 1 to estimate the spatial distribution of the recycling ratio at monthly time scales by using an iterative procedure. Spatially lumped or annual estimates can be obtained by aggregating the spatially distributed values by using precipitation as the weighting factor in the averaging process.

It should be noted that results from these semi-empirical estimations should be treated as an index for comparing the temporal (e.g. seasonal or annual) variability of recycling for a region or for comparing different regions of the globe under the assumptions made rather than be used as an absolute measure for the degree of moisture recycling over a region.

### 2.2 The NCEP datasets

The main dataset used in this study is the NCEP-NCAR reanalysis (Kalnay et al., 1996, Kistler et al., 2000) produced with the NCEP four-dimensional data assimilation model at T62 spectral resolution (2.5° lat-lon grid spacing) and 28 vertical levels. Atmospheric and surface fields including those required for the computation of the inversion characteristics and recycling ratio, such as atmospheric kinematic and moisture variables and surface precipitation and evaporation fields, are archived every 6 hours in the dataset. Evaluations of the NCEP moisture fields were carried out by Trenberth and Guillemot (1998), Janowiak et al. (1998) and Kistler et al. (2000). Since the NCEP evaporation and precipitation data have some known deficiencies over the MRB region (Kistler et al., 2000), the gridded and adjusted precipitation climatology dataset from the Meteorological Service of Canada (MSC) (Mekis and Hogg, 1999) was used instead of the NCEP precipitation fields in the calculations of the spatially or temporally lumped recycling ratios.

A comparison of the NCEP model-generated variables with independent observations and with several climatologies shows that they generally contain considerable useful information for seasonal and inter-annual variability (Kistler et al., 2000). Since this study is concerned mainly with the spatial and temporal variability of recycling over the basin, the NCEP dataset is believed adequate for the purposes of this type of study (see also the discussions in Trenberth, 1997). However, quantitative details should be treated with caution as noted earlier.
3. Results and Discussions

3.1 Recycling over the MRB

Spatial distributions of the 9-year mean seasonal precipitation recycling ratio over the MRB are given in Figure 1. Recycling over the region is typically low during the cold season (Figures 1c and d). Although the MRB is a favourable region for the development of winter cyclonic storms, which produce significant amount of precipitation over the region, surface evaporation is extremely low due to the low temperatures during these months. Precipitation from these synoptic systems is typically derived from moisture advected into the region from either the Pacific or the Arctic Ocean (Stewart et al., 1998). In addition, strong and enduring surface-based temperature inversions are common wintertime features over the MRB (Szeto, 1998), trapping most of the evaporated water vapour within the boundary layer. These factors contributed to the low degree of moisture recycling over the region during the cold season.

As expected, recycling is strongest during the warm season when both the surface evaporation and precipitation are strong over the region (Figures 1a and b). Similar to the results for the cold season, $\rho$ is largest over the eastern part of the basin, due largely to the prevailing westerly flow over the basin throughout the year. However, both the magnitude and gradients of $\rho$ are much stronger during the warm season. The 9-year (1988-1996) mean MJJ recycling ratio exceeds 0.7 over some locations of the eastern basin.
Seasonal variation of the mean monthly basin-average precipitation recycling ratio for the MRB are presented in Figure 2. These results suggest two distinct regimes in the moisture recycling over the basin corresponding to the vastly different meteorological features, surface conditions and air-land interactions over the region during the warm and cold season. The basin-average recycling ratio stays low near or below 10% between September and March. It rises rapidly during the spring and stays high throughout the summer months with the mean $\rho$ exceeding 0.3 between April and July. The strongest recycling (with the mean basin-average $\rho \sim 0.47$) occurs in June when the soil moisture storage is recharged from snowmelt and atmospheric temperature rises. Strong surface evaporation is possible over the basin under such conditions. Recycling over the region decreases rapidly again between July and September. This seasonal variability of $\rho$ is much less evident for tropical basins like the Amazon (between 2.1 for January - 2.9 for June, see Figure 9 of Eltahir and Bras, 1994). These results also suggest that the warm-season basin-average $\rho$ values for the MRB are somewhat higher than those estimated recently for the Amazon basin. Substantial annual variability in the basin-average monthly recycling ratio is however evident in Figure 2, especially during the warm season.

The 9-year mean annual recycling ratio for the basin is 0.25 (with a horizontal length scale $l \sim 1500$ km). Comparatively, estimated annual recycling ratios are ~0.25-0.35 for the Amazon basin ($l \sim 2500$ km, Eltahir and Bras, 1994), ~0.24 for the Mississippi basin ($l \sim 1400$ km, Brubaker et al., 1993), ~0.13 for the Eurasia basin ($l \sim 1300$ km, Brubaker et al., 1993) and ~0.35 for the Sahel ($l \sim 1500$ km, Brubaker et al., 1993). These values suggest that, despite the extremely inactive recycling over the basin during the cold season, the annual basin-average recycling ratio for the MRB is quite comparable to those located in the more southern or tropical
regions due to the correlation of strong precipitation and high recycling ratios for the MRB during the warm season.

Results of the recycling estimations are interpreted in terms of the atmospheric and surface settings characterizing the Mackenzie basin. The high recycling ratios and the recycling patterns characterizing the basin during the warm season are consequences of the unique topographical and climatic settings of the basin: (i) relatively low moisture influx into the region (compared to the southern basins) due to the topographic blocks over the western basin and cooler environment, (ii) active evapotranspiration over the basin during the warm season due to the extensive coverage of vegetation and water bodies, and snowmelt recharge of the soil moisture storage during spring, and (iii) precipitation is strongest over the south and southwestern parts of the basin where the juxtaposition of the strongest moisture inflow, active evapotranspiration and effective convection triggering mechanisms occurs. A significant amount of the moisture advected into the basin from the west and southwestern boundaries can be lost through the frequent convective storms that occur near the entrance region. The atmospheric moisture content is replenished through active evapotranspiration as the airflow traverses the basin. The results suggest that this locally evaporated water could contribute significantly to the precipitation over the basin, especially over the eastern parts due to the prevailing westerly flow (Figure 1).

Analysis of conditions during the anomalous precipitation years suggests that the large-scale atmospheric settings could act in concert with the basin’s unique topographic and surface characteristics to demote or to promote air-land interactions over the basin, depending on whether the basin is under the influences of a persistent large-scale high or a low pressure system (Szeto, 2002). When the region is under the influence of a persistent large-scale high pressure system (Figure 3a), the enhanced northwesterly surface flow would advect the evaporated water vapour out of the basin through the unobstructed east and southeastern boundaries, and would thus have a detrimental effect on both the precipitation and its recycling over the basin. In addition, most of the recycled precipitation would fall over the southern part of the basin where the runoff ratios are relatively low (Lawford, 1994). These effects might have contributed to the low precipitation and discharge for the basin during the spring and summer of 1995 (Stewart et al., 2002; Szeto, 2002). On the other hand, when the basin is under the influence of an enduring large-scale low-pressure system (Figure 3b), the enhanced southeasterly boundary layer flow developed under such a condition would advect both the external and locally evaporated moisture into the interior basin and towards the western topographic features. It is argued that both air-land interactions and atmospheric convection would be promoted under these situations, and resulting in enhanced precipitation and recycling activities over the basin. Moreover, much of the recycled precipitation would fall over the northwestern part of the basin where the runoff ratios are relatively high. These effects might have contributed to the record high precipitation and discharge from the basin, for example, during the summer of 1988.
Figure 3: Schematics showing the responses of precipitation and its recycling over the MRB to variations in large-scale atmospheric conditions: (a) a dry summer characterized by an enduring large-scale high pressure anomaly, (b) a wet summer characterized by an enduring low pressure anomaly, and moist southerly flows.

3.2 Recycling over the Lena River Basin (LRB)

Spatial distributions of the mean seasonal precipitation recycling ratio over the Lena River Basin are given in Figure 4. Although strong seasonality is also evident in the results, recycling is in general more active over the LRB than over the MRB throughout the year. Diagnostically, the higher recycling ratios estimated for the LRB are a result of the much weaker moisture flux over the region, when compared to the MRB. Despite the high topographic shields that exist over the upstream boundary of the MRB, magnitudes of the moisture influx into the region are relatively large compared to those for the LRB, due largely to the proximity of the MRB to the Pacific Ocean. On the other hand, both the N-S (for March-September) and E-W (for JJA) gradients of
\( \rho \) are much stronger over the MRB. The relatively uniform distribution of \( \rho \) over the LRB can be related to the relatively homogeneous horizontal distribution of evapotranspiration over the region, when compared to the MRB, which might in turn be related to the relatively more homogeneous distribution of surface cover types over the LRB. The mean annual \( \rho \) estimated for the region about 0.4.

![Figure 4](image-url)

Figure 4: As in Figure 1 but for the Lena River Basin.

Results in Figure 4 suggest that recycling is responsible for more than half of the summer precipitation (JJA) over the LRB. When averaged over the April-September period, the basin average \( \rho \) for the MRB and LRB are 0.38 and 0.48, respectively. It is of interest to note that due to the active moisture recycling over the LRB, warm-season precipitation is higher for the LRB (264 vs. 235 mm), despite both the weaker average moisture flux (315 vs. 450 mm) and evapotranspiration (430 vs. 439 mm) over the LRB. One of the reasons for the more active recycling over the LRB is the topographic shields that locate over the downstream (E and SE) boundaries of the region. These features minimize the loss of the locally evaporated moisture through the downstream basin boundary. Since such topographic shields do not exist for the MRB, a substantial amount of the locally evaporated moisture can be lost through the E and SE boundary of the basin, as evident in the frequently observed divergent moisture flux over the region during the summer months (see also Szeto, 2002).
4. Summary

Moisture recycling over the Mackenzie and Lena basins is investigated by estimating the precipitation recycling ratios over the region with the NCEP reanalysis dataset and precipitation climatology. The estimated precipitation recycling ratio for the MRB exhibit both strong seasonal and spatial variability. Recycling ratios are very low during the cold season and typically high during the warm season, reflecting the strong seasonal dependence of meteorological and surface conditions over the regions. Due to the positive correlation of strong precipitation and active recycling over the regions during the warm season, the annual average recycling ratios for the basins are comparable to those estimated previously by others for southern and tropical basins. The results suggest that recycling is responsible for close to 40% (50%) of the warm season precipitation for the MRB (LRB). Some differences in recycling activities over the two basins can be related to the differences in the physical environmental settings of the two regions. Apart from providing physical insights on how the basins’ warm season hydrologic cycle might react to variations in the large-scale atmospheric forcing, results from these analyses will also be useful, for example, in the interpretation of results from model simulations of the basins’ current and future climate. Due to the potential importance of moisture recycling in the production of warm-season precipitation over these northern basins, further investigation of the subjects is warranted when improved large-scale datasets (e.g. generated from modern high-resolution coupled regional climate models or from the next generation reanalysis products, and through field observations) are available for the regions. In addition, it is also important to estimate recycling activities over these regions by using other means that are different from the empirical numerical method employed in this study (for example, the isotope tracing method, see Sugimoto, 2001, this volume) for the purposes of cross-validation and to provide further insights into the subject.

5. References


LAND SURFACE STUDIES
Present-day Landscape Dynamics and Permafrost Variation in Yakutia

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1. Introduction

Permafrost covers about one-quarter of the Earth’s surface and more than one-half of the Russian territory. Distribution, thickness and temperature of perennially frozen ground are predetermined by both the last epoch’s and present-day landscape-climatic conditions.

Cryogenic processes form such permafrost landscapes as alases, bylars (polygonal network of mounds and trenches formed by the melting of upper portions of wedge ice), pingos, low-center polygons in excessively wet areas, sorted polygons (resulting from frost sorting) and others. Permafrost landscapes also include complexes whose lithogenic base consists of frozen bedrock.

Permafrost landscapes, as are all units of natural environments, are dynamic systems. They vary all the time both under anthropogenic factors and natural processes (primarily climatic).

This paper focuses on present-day dynamics of the environment and variations in the upper permafrost in Yakutia. The emphasis is on the dynamics of landscapes and permafrost after anthropogenic impact and present-day climate changes.

2. Results

Variations in surface conditions have a rather strong impact on the development of permafrost. This is corroborated by the active dynamics of ground temperature and active layer thickness as well as related surface variations.

In this paper we would like to call your attention to some aspects of the influence of surface condition variation on permafrost.

2.1 Anthropogenic Successions

The vegetation dynamics of post-disturbance impact have a great influence on the development of permafrost landscapes. Anthropogenic successions connected with vegetation restoration on disturbed sites make a practice of having general regulations for development. At present, tree cutting and fires disturb large areas. In the vicinity of Yakutsk they occupy about 70-80% of the territory. Our research shows that the influence of vegetation restoration on permafrost conditions is quite considerable. Based on the study of post-disturbance landscape dynamics, we constructed a succession model representing regularities in environmental restoration in Central Yakutia (Fedorov, 1996) (Figure 1). Investigations of changes in permafrost conditions in the successions of varying age allowed us to understand the restoration patterns of permafrost.
conditions, including soil temperature and thicknesses of the active layer and the protective layer. Figure 1 shows the deviations of primary natural conditions.

Figure 1: A succession scheme of landscape evolution at Umaibyt site – (a), variability of ground temperature (Kt), seasonal thaw depth (Kζ) – (b) and protective thickness (Kp) – (c).
Our results suggest that most landscapes have a tendency toward invariant conditions. At the Umaibyt site, for example, a cooling trend in soil temperature up to 0.5°C has been observed over the period 1980-2000 in accordance with the regularities in succession development. The succession-based method of permafrost research is now being refined, and retrospective and predictive permafrost-landscape maps of the study area are under preparation.

Phytocenosis productivity in some tundra landscapes increases by 15-20 times after the disturbance, becoming a stabilizing factor for permafrost conditions and the whole landscape.

### 2.2 Cryoecological Stress in Landscapes

Variation in permafrost conditions depends mainly on variation in climate. Whether destructive processes will occur or the natural environment will be restored depend on the state of the natural environment at the time of disturbance. Therefore it is necessary to study conditions of formation of cryoecological stress in landscapes of Yakutia. For its estimation we used variations in series of meteorological parameters, and geocryological and dendrochronological data. These data are characterized by definite cyclic changes (Fedorov, 1996). Stress situations in permafrost development are formed cyclically, and in Central Yakutia this happens at 11-12 year intervals (Figure 2).

**Figure 2:** Variability of cryoecological stress in landscapes of Central Yakutia.

The generalized coefficient was calculated based on normalized anomalies of permafrost temperature, thawing index, active layer thickness, summer precipitation index and birch ring growth index for each year. Time-series analysis revealed some regularity of the landscape dynamics due to all these factors. Our estimates show that during the period considered there were four cryoecologically stressed periods in Central Yakutia. Statistical analysis of the generalized coefficient indicates that the highest correlations are with active layer thickness ($r^2=0.68$, $n=55$).
2.3 Regional Peculiarities of Landscape Dynamics

In the landscape evolution study, consideration should also be given to the *regional distribution of dynamic characteristics*. We (Fedorov and Svinoboev, 2000) compiled schematic maps of trend value distribution for mean annual, mean January and mean July temperatures in Yakutia (Figure 3) for the purpose of landscape dynamics research. Data from 95 weather stations over the period 1951-1989 were used. This approach gives an indication that the present-day landscape dynamics may considerably vary from region to region.

**Figure 3:** Spatial distribution of a linear trend in (a) mean annual air temperature, (b) mean January and (c) mean July temperature for the period 1951-1989.
Figure 3a shows the maximum positive trend of mean annual temperature (0.03 to 0.04°C per year) in Central Yakutia and Oymyakon-Verkhoyansk region. A region without a trend is indicated along the area of tundra and near-tundra forests, and a negative trend (averaging -0.01 to -0.02°C per year) is indicated in Anabar tundra, on the Novosibirsk islands and in Kolyma-Indigirka tundra. Generally, about 75% of the territory of Yakutia lies in the region of positive trends for mean annual air temperature.

The schematic map for January trends (Figure 3b) depicts a clear temperature decrease in the northwest of Yakutia in the interfluves of the Anabar-Olenek-Lena Rivers and in the estuary of the Aldan River. The rest of the territory has an increased temperature. The maximum increase in January is indicated in Central Yakutia (0.07°C per year).

In July, the area of tundra and near-tundra forests differs in temperature with a decrease averaging 0.01-0.03°C per year (Figure 3b). The southern and southwestern parts of Yakutia are characterized by temperature decrease and among these are Prilensky and Midviluysky regions. The area between these regions is characterized by a positive trend averaging 0.01°C per year. Maximum temperature increase in July is observed in the upper part of the Yana river basin.

Many authors, e.g. Balobaev (1997) in Yakutia and Varlamov et al. (1998) in East Siberia and Far East, point to the sign differences in the trends.

Research of such problems gives the knowledge of the general directionality of permafrost development for definite periods of time. This is very important for determining the stability and for prediction.

2.4 Landscape Subsidence

One of the characteristic features of present-day landscapes in the North is subsidence connected with cryogenic processes. The greatest landscape changes occur with thermokarst development. Observations for surface dynamics of post-disturbance impact were conducted at Yukechi and Kys-Alas sites near Yakutsk (Fedorov and Fukuda, 2001). These sites are characterized by a great amount of ground ice in surface deposits. Measurements were made on undisturbed sites at reference points grounded in permafrost at the depth of 4 m. Subsidence on tussock sites at small thermokarst depressions was 5-10 cm y⁻¹ (Figure 4). The maximum subsidence (17-24 cm y⁻¹) is typical for depressions with initial thermokarst lakes. However, surface subsidence is indicated even on well-drained surfaces of inter-alas areas (Figure 5).

At the Kys-Alas site, inside of the GAME-Siberia study area (Spasskaya Pad) we conducted observations for subsidence on young clear cuttings of 1990 and 1996. The surface subsidence here was 2-5 cm y⁻¹.
Figure 4: Surface subsidence on the Yukechi location: a) Site 2. C: marker, undisturbed inter-alas terrain; D – incipient thaw depression; 1 and 2: centers of polygons within thaw depression, and b) Site 5. A: marker, undisturbed inter-alas terrain; 2 and 3: centers of polygons within thaw depression, the marker No.3 is below the water edge since 1994; 6 – incipient thaw depression.

Figure 5: Surface subsidence on the Yukechi location (well-drained inter-alas surfaces). Markers A, C, D and ad68 – Site 2; markers B and bc28 – Site 3.
3. Conclusion

We have discussed the effect of variations in surface conditions on the permafrost landscapes. There are anthropogenic landscapes whose development differs from general regulations. Therefore, when studying present-day climate effect on permafrost, the vegetation successions must be taken into account. The precise identification of anthropogenic succession age allows the proper understanding of the permafrost state for a definite period of time.

The state and the future of permafrost landscapes depend heavily on cryoecological conditions at the period of anthropogenic impacts. Greatest changes in permafrost landscapes may occur during the period of cryoecological stress formation.

The landscape and permafrost dynamics have regional peculiarities. When there is climate warming and permafrost subsidence in one region, the reverse effect is possible in another one. Therefore, when studying landscapes and permafrost, the regional peculiarities of dynamic characteristics of nature must be taken into account.

The cryogenic processes activation is the indicator of disturbing processes in cryolithozone. Therefore, its characteristics clearly point to the tendency for nature development at a definite period.

4. References


1. Introduction

As a part of GAME-Siberia, hydrological and meteorological observations for a basin water/energy balance have been carried out in Siberian tundra region near Tiksi, Sakha Republic, Russian Federation since 1997. The tundra area has a unique water cycle due to the existence of frozen ground, drifting snow and tundra vegetation. The frozen ground limits subsurface flow, and subsurface storage and its change is small. Strong wind and low vegetation results in redistribution of snowcover and formation of snow drifts leeward of ridges and in depressions. Melt water from the snowdrifts becomes a source of summer runoff for the tundra watershed, where liquid precipitation is small. Low vegetation such as lichen, moss and sedge, play an important role in evaporation over the tundra watershed, and soil water content in turn determines the vegetation distribution. These energy/water fluxes are also dependent on the atmospheric conditions, which are mainly characterized by wind direction. In this paper, the annual water balance and seasonal variation of heat balance at the tundra surface are reported.

2. Study Area

A 5.5 km² experimental watershed, located 7 km south of Tiksi, Sakha Republic, Russia has been selected for this study (71°N, 129°E). It is a tributary of the Suonannav River and the elevations range from 40 to 360 m.a.m.s.l. We have initiated monitoring of snowmelt and summer heat and water balances at patch and basin scales, installation of a fully instrumented meteorological station, monitoring the variation in soil moisture, evaluating the variation of one-dimensional vapour and energy fluxes, and monitoring variation in the active layer (thickness, temperature, moisture) over a 1 km² grid.

Climatic data have been collected at the nearby Polyarka Hydrometeorological Station since 1932. The mean annual air temperature (30-year mean from 1955 to 1984) is -13.5°C. The maximum and minimum air temperatures recorded during this period were 32.6°C and -53.6°C, respectively. Mean annual wind speed is 5.0 m s⁻¹, with the prevailing wind direction being northeast in summer and southwest in winter. Mean annual precipitation is 345 mm.
3. Heat Balance

Micro-meteorological observations have been carried out in the flat plain of the tundra region using a 10 m instrumentation tower, and soil temperature and water content in the active layer were measured. In order to estimate the heat balance components, the heat balance equation (Eq. 1) was solved for the surface temperature, $T_s$, by iteration:

$$S \downarrow - S \uparrow + L \downarrow - \sigma T_s^4 + H_s + LE + G = 0$$  \hspace{1cm} (1)

$$Rn = S \downarrow - S \uparrow + L \downarrow - \sigma T_s^4$$  \hspace{1cm} (2)

$$H_s = \rho C_p D(T_a - T_s)$$  \hspace{1cm} (3)

$$LE = 0.622 \beta \rho L_v D(e_a - e_{ss}) / P$$  \hspace{1cm} (4)

$$G = K(T_s - T_z) / z = 0$$  \hspace{1cm} (5)

where $S \downarrow$ and $S \uparrow$ are the incoming and outgoing short wave radiation, respectively, $L \downarrow$ the incoming long wave radiation, $\sigma$ the Stefan-Boltzman constant, $T_s$ the surface temperature, $H_s$ the sensible heat flux, $LE$ the latent heat flux, $G$ the heat flux in soil, $\rho$ the density of air, $C_p$ the specific heat capacity of air, $D$ the turbulent transfer coefficient, $T_a$ the air temperature, $\beta$ the evaporation efficiency, $L_v$ the latent heat of vaporization, $e_a$ the vapour pressure of air, $e_{ss}$ the saturation vapour pressure at the surface temperature $T_s$, $P$ the barometric pressure, $K$ the heat conductivity of soil and $T_z$ the soil temperature at the depth of $z$. The turbulent transfer coefficient $D$ is changeable with the stability of the atmosphere as shown by Thom (1975). The heat conductivity of soil is changed linearly with the soil moisture at the depth of 0.05 m. To solve Eq. 1, first an arbitrary surface temperature is given, then the transfer coefficient is calculated, and the surface temperature is obtained iteratively by the Newton-Lapson method. Using the newly obtained surface temperature, a new transfer coefficient is calculated and the same method is repeated until the surface temperature change becomes negligibly small. When the surface temperature is obtained, the heat balance components are calculated using Eqs. 2-5.

The results of the daily values of the heat balance components for 1998 and 1999 are shown in Figure 1. The net radiation reaches a maximum near the end of June and then gradually decreases. The heat flux in soil is the smallest of the components and does not change much. The sensible heat flux is large at the end of June, does not exhibit a clear seasonal variation, and sometimes shows positive values. This coincides with the large latent heat flux. These are all related to wind direction, which will be discussed later.

The 10-day mean heat balance components and their ratio to the net radiation for the period from June 10 to September 10 in 1998 and 1999 are shown in Figure 2. The net radiation was greatest at the end of June for both years. The latent heat flux is larger than the sensible heat flux during most of the periods. The heat flux in soil is the smallest among the components and decreases gradually towards fall. Net radiation is partitioned into sensible, latent and soil conductive heat fluxes and the ratios of these components to the net radiation are shown in the bottom of
Figure 2. The mean ratios are 25-30% for sensible heat flux, 50-55% for latent heat flux and 20% for soil conductive heat flux.

Figure 1: Daily heat balance in 1998 and 1999.

Figure 2: 10-day mean heat balance components and the ratios to the net radiation for 1998 and 1999.
4. Wind Direction Dependency

We have seen positive values of sensible heat flux and large latent heat flux at the same time. This is closely connected to the wind direction. Figure 3 shows the wind directional dependency of the heat balance and meteorological components in 1998. From the top and left to right, frequency of wind direction; ratios of net radiation, sensible heat flux, latent heat flux, soil conductive heat flux; and average air temperature and vapour deficit are shown. The “ratio” refers to the ratio of total heat flux in the specific wind direction to the total heat flux for all wind directions. The most frequent wind direction is from the east and northeast, which are the same as the climatic prevailing wind direction at Poliarka Hydro-meteorological station in summer. Easterly and northeasterly winds are onshore winds, and, conversely, southwesterly and westerly winds are offshore winds. The sensible heat flux is large and positive for onshore winds but mostly small or negative for offshore winds. This is obvious for most of the periods, especially in July (July 3). The latent and soil conductive heat fluxes do not clearly depend on the wind direction. For offshore winds, air temperature is high and the vapour deficit is large, whereas onshore winds are cold and have smaller vapour deficits. The wind direction dependency of the heat balance components was also found in Alaska and Northern Canada (Yoshimoto et al. 1996, Rouse et al. 1987, Rouse 1984).

5. Summer Water Balance

Snowdrifts are common features in the tundra region due to strong winds and low vegetation such as mosses, sedges and lichens. It is usually formed leeward of hills and ridges, and in depressions such as streambeds (Liston et al., 1998). Since precipitation is small in the Arctic, meltwater from snowdrifts, which remain as snow patches until the middle of summer, is one of the main sources of summer runoff. In this study the summer water balance of a tundra watershed was examined considering the amount of meltwater from the snow patches.

The water balance of a watershed is expressed by the following equation:

\[ P + M = E + Q + dS \]  

(6)

where \( P \) is the rainfall amount, \( M \) the snowmelt amount from snow patches, \( E \) the evapotranspiration, \( Q \) the discharge and \( dS \) the storage change in soils. All variables are basin averages. \( P \) and \( Q \) were measured in the watershed. \( M \) can be expressed by \( M = ma \), where \( m \) is the mean melting intensity of snow patches and \( a \) is the ratio of snow patch area to the whole watershed area. \( M'm' \) is estimated by a degree-day method.
Figure 3: Wind directional dependency of heat balance and meteorological components.
Mean basin evapotranspiration was obtained assuming that: a) there are 2 types of ground surface: moss-sedges and gravel, b) snow only exists on gravel, c) the evapotranspiration over gravel is linearly correlated with evapotranspiration over moss-sedge, and d) there is no net evaporation/condensation from snow. E can be expressed as:

\[
E = S_w E_w + (S_d - a)E_d = S_w E_w + (S_d - a) f_d E_w
\]  \hspace{1cm} (7)

where \(E_w\) and \(E_d\) are the evapotranspiration over moss-sedge and gravel, respectively, and \(S_w\) and \(S_d\) are the ratio of moss-sedge area and gravel area, respectively, and \(f_d = E_d / E_w\).

The storage term in the water balance equation was assumed to be the integrated virtual discharge extrapolated after the local minimum discharge point with a constant recession coefficient, and was determined by linearly fitting the recession part of hydrograph plotted on the logarithmic axis. The storage change was obtained as the difference in the storage values calculated at the two local minimum discharge points. In the water balance equation, the ratio ‘a’ of snow patch area to the watershed area was a tuning parameter and therefore determined so that all the components were balanced. After the snow patches disappeared, the residual \(X\) was introduced to make the components be balanced.

Seasonal variation of water balance components in 1997, 1998 and 1999 are shown in Figure 4. Precipitation was smaller at the end of July, and relatively larger at the beginning of July and the middle of August every year. The snowmelt amount decreased with decrease in snow patch area in the 3 summer seasons; however, for over a month it affected the water balance in the watershed after a major snowmelt period. Discharge is the major component in the outputs, but at the end of July when inputs were small, the evapotranspiration was larger than the discharge. The mean evapotranspiration was small, 1-2 mm per day. This result agrees with observations in the Alaskan Arctic (e.g. Kane et al., 1990). The storage change was smaller than the other components, ranging from -0.6 to +0.8 mm per day, and it did not change seasonally.

The yearly changes of the precipitation, snowmelt amount and discharge were large, whereas the yearly change of the evapotranspiration was small (not shown). The ratio of discharge to total input amount \((Q/(P+M))\) was larger than the same ratio of the other output components. \(Q/(P+M)\) values during the snowmelt period were 0.77 in 1997, 0.68 in 1998 and 0.27 in 1999; and during the snow-free period, \(Q/P\) \((M=0)\) values were 0.58 in 1997, 0.75 in 1998, 0.66 in 1999. For the whole period, \(Q/(P+M)\) was 0.75 in 1997, 0.59 in 1998 and 0.69 in 1999. These results are consistent with observations in Alaskan and Canadian Arctic, where \(Q/P\), called the flow rate, is typically from 0.7 to 0.8 for rainfall events (Anderson, 1974; Findley, 1969; Kane and Carlson, 1973). Since there are no snow patches in their watershed, the total input is \(P\), whereas ours is \(P+M\). The ratio of the evapotranspiration to the total input was 0.24 in 1997, 0.36 in 1998 and 0.64 in 1999, and yearly change of this ratio is largely due to large variation in the total input, not in the evapotranspiration. The ratio of meltwater from snow patches to the total input was 0.61 in 1997, 0.65 in 1998 and 0.07 in 1999. Except in 1999, the contribution of snowmelt water to the water balance was large. If the period analyzed was limited to the period when snow patches exist in the watershed, the ratio became larger, 0.69 in 1997, 0.71 in 1998 and 0.42 in 1999.
Figure 4: Seasonal variation of water balance of the watershed during summer in 1997, 1998 and 1999. (a) is for the input (precipitation and snowmelt) and (b) is for the outputs (discharge, evapotranspiration, storage change and residual).

6. Summary

In order to better understand the water cycle over tundra, micro-meteorological and hydrological observations have been carried out near Tiksi, Eastern Siberia and seasonal variation of energy budget components and characteristics of the summer water balance were investigated.

The heat budget equation at the tundra surface was solved approximating the surface temperature by iteration. The seasonal changes of heat balance components were obtained for 1998 and 1999. Net radiation is partitioned to sensible, latent and soil conductive heat fluxes and the ratios of these components to the net radiation are: 25-30% for sensible heat flux, 50-55% for latent heat flux and 20% for soil conductive heat flux. The energy budget over tundra changed with wind direction. The southwesterly wind was warm and dry, making the sensible heat flux small or towards the ground surface, and the northeasterly wind was cold, giving the large sensible heat flux to the atmosphere from the tundra surface.
Due to strong wind in winter and low vegetation height, snowdrifts are a common feature in tundra area, and they remain as snow patches until the middle of summer, supplying meltwater to river flow. In the liquid water balance analysis, we introduced the ratio of the snow patch area to the whole watershed area as a tuning parameter for taking into account the meltwater from the snow patches. The results of this study can be summarized as follows:

1. The contribution of meltwater from snow patches to discharge continued in summer;
2. Rainfall and snowmelt amount changed remarkably from year to year;
3. The major output component was discharge;
4. Contribution of evapotranspiration to the output changed from year to year; and
5. Change in storage was small when compared with the other components.

7. Acknowledgments

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8. References


Cold Region Lakes and Landscape Evaporation

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1. Introduction

The goal of this paper is to present a preliminary evaluation of the role of lakes in landscape evaporation. Lakes of various sizes are ubiquitous in high latitude subarctic and tundra regions. For example, in the central Mackenzie River Valley of northwestern Canada, there are about 32,370 lakes whose frequency-size distribution is shown in Figure 1. For the whole Mackenzie River Basin, at least 8.5% of the total surface area (1.8 x 10⁶ km²) is covered in lakes equal to or larger than 1 km² but this number is probably underestimated (Normand Bussieres, personal communication). Also there are many lakes and ponds with areas < 1 km² (Figure 1).

![Figure 1: Frequency-size distribution of lakes in the central Mackenzie River Basin (Lower Slave District).](image)

With such abundance, lakes are important features in regional climatic, meteorological and biogeochemical cycling. The purpose of this paper is to examine the potential role of lakes in the regional surface energy balance, and to link this to the frequency-size distribution of lakes. Toward this end we will utilize data from high latitude lakes of various sizes, will characterize the magnitudes and temporal behaviour of evaporation, and will examine the impacts of combinations of various-sized lakes and land-lake distributions on regional energy balances during the ice-free period.
For the purposes of this paper, lakes are classified as small (<1 km²), medium-sized (1-50 km²), and large (>50 km²). The lake data that have been utilized in this study are shown in Table 1.


<table>
<thead>
<tr>
<th>Measurement Sites</th>
<th>Latitude</th>
<th>Longitude</th>
<th>EB Methods</th>
<th>Lake Size (km²)</th>
<th>Mean Lake Depth (m)</th>
<th>Lake Class</th>
<th>Ref</th>
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<td>63º 35' N</td>
<td>113º 54' W</td>
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<td>--</td>
<td>--</td>
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<td>114º 22' W</td>
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<td>0.6</td>
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<td>2</td>
</tr>
<tr>
<td>Golf Lake</td>
<td>58º 40' N</td>
<td>94º 40' W</td>
<td>BREB</td>
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<td>0.9</td>
<td>small</td>
<td>2, 4</td>
</tr>
<tr>
<td>Skeeter Lake</td>
<td>63º 35' N</td>
<td>113º 54' W</td>
<td>BREB</td>
<td>0.05</td>
<td>3.2</td>
<td>small</td>
<td>2</td>
</tr>
<tr>
<td>Sleepy Dragon Lake</td>
<td>62º 55' N</td>
<td>112º 55' W</td>
<td>BREB</td>
<td>5.50</td>
<td>12</td>
<td>medium</td>
<td>2</td>
</tr>
<tr>
<td>Great Slave Lake</td>
<td>63º 35' N</td>
<td>113º 54' W</td>
<td>EC</td>
<td>18,500</td>
<td>60</td>
<td>large</td>
<td>2, 3</td>
</tr>
</tbody>
</table>

The upland shield measurement site, located in the Yellowknife River Basin, has vegetation typical of the subarctic Canadian Shield (Spence and Rouse, 2002) with stands of black spruce (26%), mixed stands of spruce and aspen (26%), a peat wetland (20%) and exposed bedrock (28%). The open forests have an understory dominated by dwarf birch, Labrador tea, blueberry and other shrubs. The soils in the wetland patches include a sphagnum moss mat up to 30 cm deep above thick organic soils. Exposed and lichen-covered bedrock is scattered throughout the footprint area of measurement. Evaporation magnitudes and cycles depend on factors of frozen soils, organic soils, ponding in bedrock reservoirs, and vegetation type and phenology as well as on the radiation balance and atmospheric temperatures. Spence and Rouse (2002) show that it can undergo large variation from year to year. On average the evaporation cycle appears to start in mid-March, reach a peak in May and June, and end in mid-September to cover a period approaching 6 months.

For the smaller, shallower lakes, as snow disappears in late winter and the ice-cover thins, solar radiation can penetrate the water beneath and with final thaw, solar heating of the shallow lakes raises their temperatures rapidly. This allows vigorous evaporation to start early in the thaw season. Subfreezing temperatures in early winter coincide with short daylight periods and freeze-up can occur quickly. The initial formation of ice puts an effective lid on outgoing turbulent heat fluxes and the evaporation cycle ends. In subarctic latitudes the period of evaporative heat loss from shallow lakes is in the order of 4 months (Rouse et al., 2000) and this becomes less in northern tundra regions.

For large deep lakes such as Great Slave (GSL), the thermal exchange cycle in 1998 shows a broad winter minimum, a spring heating period, a summer period of maximum heat storage, and a fall-winter cooling phase. During spring, the total heat flux is dominated by high net radiation
contributing to lake heating (Schertzer et al., 2000). Lake heating proceeds slowly because of the large vertical mass and deep thermal mixing particularly during storms. A complete vertical mass exchange of surface and deep water occurs at 4°C after which the development of thermal stratification proceeds. Evaporative heat loss into the atmosphere often will not commence until mid-July after which the entire lake is freely evaporating to the overlying atmosphere. There is a strong asymmetry between heating and cooling rates due to large heat storage in spring and summer and increased evaporation during the fall ice-free period as the warmer water surface exchanges both heat and mass with the cold overlying air. As in the shallow lakes, a complete winter ice and snow cover effectively restricts air-water heat and mass exchange. Evaporation from GSL operates over a period ranging from 5 to 6 months.

As might be expected, evaporation from medium-size lakes has characteristics of both the small and the very large. They heat more rapidly in the spring than the very large lakes and cool more rapidly in the early winter and this is reflected in their evaporation cycles. Sleepy Dragon Lake has an evaporation period in the order of 5 to 6 months.

2. Magnitude and Temporal Patterns of Evaporation

As noted in Eaton et al. (2001), lakes of all sizes have the largest annual evaporation rates of any high latitude surfaces. With reference to Figure 2, if evaporation amount from upland shield terrain (land) is taken as unity then ratios of (land:small lakes:medium lakes:large lakes) gives (1:1.28:1.65:2.11) based on our measured data.

![Annual Evaporation](image)

Figure 2: Average annual evaporation totals from land and from lakes of various sizes.
Temporal patterns of the evaporation cycle are highly distinctive for different-sized lakes (Figure 3). Although the subarctic shield terrain is the earliest to begin its evaporation cycle in spring, it undergoes strong moisture limitations during July and August (Spence and Rouse, 2002) and ends its evaporation cycle in September due to short days and cold temperatures. Shallow lakes begin evaporating close to a month later than nearby land surfaces and their peak evaporation rates during June, July and August are substantially larger. Also, they cease evaporating somewhat later than the bedrock dominated terrain. Moderate-sized lakes as represented by Sleepy Dragon begin evaporating a month later than their small counterparts but, in compensation, their cycle lasts substantially longer into late fall-early winter. Great Slave Lake shows a further 1-month lag in the onset of spring evaporation but reaches its largest magnitudes of evaporative heat loss well into the winter period and continues to evaporate up to three months later than the land surface.

Figure 3: Temporal patterns of mean monthly evaporation for different-sized lakes.

3. Landscape Evaporation and its Implications

By combining the data in Figures 1 and 3, one can derive a picture of the influence of lakes on landscape evaporation (Figure 4) as it applies in the central Mackenzie Basin. As well as the difference in total magnitudes due to the presence of lakes, the salient features of the seasonal patterns are a dampening of evaporation rates in early summer, and the long extension of the evaporation period into the winter season.
Figure 4: Regional patterns of actual landscape evaporation compared with the probable pattern of a lake-free (all land) landscape (Lower Slave District).

The implications of the patterns in Figure 4 are many and only a few will be noted here. On an annual basis about 23% of the precipitation falling in the Mackenzie River Basin has its origins in evaporation from the basin itself (Szeto, 2001). This precipitation recycling is highest in the eastern part of the basin where the majority of lakes and wetlands are concentrated. There is a peak in June at about 50%. During winter, the precipitation recycling remains close to 10%. It is evident from Figure 4 that an important component of the evaporative input is due to lakes of all sizes within the basin.

Evaporation involves, of course, the transfer and release of large amounts of latent heat. Since the temporal patterns in evaporative heat loss from lakes are echoed by the sensible heat losses, the total represents a large heat release into the atmosphere during the early winter period which can ameliorate the temperature especially downwind of the larger lakes. It also is a source of moisture to increase the snowfall downwind of the larger lakes. In the spring and early summer, the large lakes can foster persistent inversions accompanied by low level cloud (fog) so that daily temperatures can be 5°C or more colder than nearby terrestrial areas. This can reduce locally induced rainfall due to the dampened buoyancy of the boundary layer.

Inevitably the impact of climate change on the lakes must be addressed. Preliminary evidence suggests that even large lakes such as Great Slave Lake can respond rapidly to warmer atmospheric conditions as seen in 1998 (Figure 2). The smallest lakes can double their seasonal evaporation totals under warm and sunny conditions (Rouse et al., 2000). The water balance of lakes becomes an important issue as the climate warms. With strongly increased evaporation, unless the precipitation increases an equivalent amount, shallow lakes will dry up and deeper lakes will be reduced in volume. Reduced evaporation will feed back to the precipitation cycle.
and could reduce the precipitation-recycling ratio in the Mackenzie Basin in a positive feedback scenario. The net effect would be to reduce the landscape evaporation, change the temporal patterns, and reduce the contrast between evaporation from terrestrial and aquatic environments. Thermal structures are important to all limnological responses of high latitude lakes (Rouse et al., 1997). Warmer temperatures, longer ice-free periods and potentially lesser lake volumes will influence floral and faunal characteristics and carry right through to the dissolved organic carbon content that influences the penetration of solar radiation into lake waters.

4. References


1. Introduction

A major focus of this investigation is to understand the role of lakes in the surface climates of cold regions. This is accomplished through use of detailed field measurements, use of remote sensing SSM/I and AVHRR techniques, and application of lake models. This report provides a brief synopsis of the current research conducted on Great Slave Lake with an emphasis on the annual temperature cycle, with a brief description of the heat exchange, evaporation and heat storage research. Important advances in knowledge of the lake responses have occurred through the integration of research between lake measurement, modelling and remote sensing approaches.

2. Distribution of Lakes in the Mackenzie Basin

The Mackenzie Basin (1.8 x 10^6 km^2) has a very large number of lakes (i.e. 30,000+) of various sizes (Rouse et al., this proc.). Figure 1a shows a distribution of lakes based on AVHRR imagery (Bussières, 2001) which illustrates that the largest concentration of water bodies in the MAGS Basin is located in the eastern half of the basin. Based on AVHRR, Bussières (2001) has estimated that there are 8900 lakes greater than 1 km^2 with a total surface area of 1.5 x 10^5 km^2. In the MAGS basin there are several large lakes, for example, Great Bear Lake (31,328 km^2), Great Slave Lake (28,568 km^2), Lake Athabasca (7,935 km^2), Lake La Martre (1,776 km^2), and Lesser Slave Lake (1,168 km^2). Much of the current research on the large lake component of MAGS is focused on Great Slave Lake.

3. Latitudinal Variations: Temperature and Evaporation

There are few detailed observations of heat or mass fluxes from the vast number of lakes in the MAGS Basin. Rouse et al. (this proc.) has developed a conceptual model of landscape evaporation response based on research within a study site 61°N to 65°N (1 x 10^5 km^2). In this region, lakes represent about 35% of the landscape surface area and include lake distributions categorized as small lakes (0-1 km^2, 95.9% of lakes), medium lakes (1-50 km^2, 4.1% of lakes) and large lakes (>50 km^2, <0.1% of lakes). The model suggests that within a particular landscape, quite different temporal responses in evaporation can be expected not only from land
surfaces but also between lakes sizes - largely related to heat storage capacity. More research needs to be conducted to quantify this relation over the large latitudinal expanse of the MAGS Basin.

Figure 1: (a) Distribution of lakes within the Mackenzie Basin based on AVHRR 1 km resolution; (b) Surface water temperature distribution for selected large deep MAGS lakes based on 1999 AVHRR data (based on Bussières, 2001).

Water temperature is one of the most important variables for determining the response of the lake to heat and mass exchanges. It is impossible to instrument all the lakes in the basin; consequently, reliance on remote sensing techniques is essential. Bussières (2001) has derived seasonal temperature curves for a subset of lakes based on AVHRR (Figure 1b) which show good correspondence with lake observations. An essential feature of the seasonal curves in Figure 1b is the latitudinal variability in timing of water temperature increases and temperature amplitudes for various lakes especially in the spring period. Continued collaboration between remote sensing scientists and lake investigators is essential to refine techniques for inclusion of weather fluctuations into the AVHRR derived temperature curves.

4. Bathymetry of Great Slave Lake

Significant research effort has been focussed on the surface processes of Great Slave Lake as an example of a large deep high latitude lake (e.g. Blanken et al., 2000; Rouse et al., 2000b; Schertzer et al., 2000a). The bathymetry of the lake and lake measurement sites are shown in Figure 2a. The main-lake has a maximum depth of 187 m, a surface area of $1.85 \times 10^{10} \text{ m}^2$, and a total volume of $5.96 \times 10^{11} \text{ m}^3$. 


5. Annual Temperature Cycle

5.1 Annual Water Temperature Cycle

An example of the annual water temperature cycle is given in Figure 2b for Great Slave Lake for the period mid-June 1998 and extending to mid-June in 1999.

5.2 Winter Temperature Regime

Great Slave Lake becomes isothermal at about 9°C in mid-October (Figure 2b). As a consequence of radiative and turbulent heat losses and deep vertical mixing resulting from high winds during the fall period (Schertzer et al., 2000a), lake temperatures fall below the temperature of maximum density (4°C) in mid-November and reach minimums (~0.1°C) that extend from December to April. Spring overturn can be expected to occur (at 4°C) at about mid-June.

Great Slave Lake has complete ice cover for approximately 6 months each year. Figure 3 shows historical ice break-up and freeze-up periods on Great Slave Lake derived from SSM/I (Walker et al., 1999). Under average conditions, the main body of the lake is expected to be ice-free by ~June 18 (DOY=169) and ice freeze-over is expected ~December 7 (DOY=341) with an average ice-free period of 172 days. Considerable variability in the break-up and freeze-up dates occurs. Over the period 1988-2000, the longest ice-free period (213 days) occurred in 1998 associated with an intense El Nino. Figure 4 shows a sequence of ice conditions on Great Slave Lake during 1998 based on SSM/I passive microwave (85 GHz) satellite data. Extensive open water areas are evident by May 27, which is about 3 weeks earlier than the average. Accurate determinations of the length of the ice-free period are critical for modelling of the lake heat and mass exchange and heat content cycle.
5.3 Summer Temperature Regime and Heat Fluxes

Year-round thermal observations on Great Slave Lake have provided critical information on the seasonal temperature cycle (Figure 2b). Great Slave Lake is a dimictic lake, meaning that the water column is completely mixed vertically once in the spring and once in the fall. The overturn events mark the beginning of the summer thermal stratified season.

Satellite remote sensing using SSM/I (Walker et al., 1999) has shown utility in detecting the start and end of the spring ice break-up period (Figure 5a). Bussières (2002) and Bussières et al. (2002) combined field temperature observations with AVHRR (Figure 5b) to detect when the lake is completely ice-free (i.e. >4°C). Both SSM/I and AVHRR techniques have shown good correspondence with lake observations.
Figure 5: (a) Spring ice break-up (SSM/I; Walker et al., 1999) and (b) surface water temperature (AVHRR; Bussières, 2000) compared with observations at the ODAS site (e.g. Schertzer et al., 2000a) for Great Slave Lake in 1999.

Intensive observations of lake meteorology and thermal structure have been conducted on Great Slave Lake during the summer field programs (e.g. Schertzer et al., 2000a). Observations show that the lake temperature structure can be strongly influenced by strong winds which can cause large wave heights and deepening of the upper vertically mixed layer. Preliminary modelling of station heat flux components (Figure 6) show that sensible heat flux is a small component of the total exchange while larger evaporative fluxes in the fall months become an important part in the lake heat losses.

Figure 6: Heat flux components at the ODAS station in Great Slave Lake in 1998. (Schertzer et al., 2000a)
While detailed measurements are being conducted on solar and longwave radiation fluxes on Great Slave Lake associated with the GEWEX-MAGS investigations, there are no such measurements on other large deep lakes in the basin. An alternative is to derive flux estimates based on satellite remote sensing. Leighton and Feng (2000) have provided a detailed description of the determination of net surface solar radiation (NSSR) derived from AVHRR observations. Top of the atmosphere fluxes (TOA) were derived from AVHRR using a narrowband to broadband conversion algorithm. NSSR was derived from TOA fluxes using the algorithm of Li et al. (1993). Figure 7 illustrates the correspondence between NSSR from NOAA-12 satellite data compared with observations derived from the lake buoys. The mean difference between satellite derived NSSR and buoy observations is 5.5 Wm$^{-2}$ with a standard deviation of 52.4 Wm$^{-2}$. Advances in the remote sensing derivation of NSSR are critical for modelling of lake thermodynamical and hydrodynamical processes in lakes especially in the northern region.

![Comparison of net surface solar radiation (NSSR) derived from AVHRR with buoy measurements in Great Slave Lake (based on Leighton and Feng, 2000).](image)

6. Interannual Variability of Heat and Mass Transfers

Understanding the interannual variability in heat and mass transfer components of lakes within the northern climate system is an important area of research in order to reduce the uncertainty in understanding the impacts of climate changes. For Great Slave Lake, intensive measurements conducted during the CAGES years 1998 and 1999 have shown that the anomalously warm year in 1998 (El Nino year) resulted in a longer ice-free season compared to other years over the period 1988-2000 (Figure 3). This response contributed to a higher cumulative evaporation (Figure 8a) and heat content (Figure 8b) in 1998 compared to other years, indicating that the northern lakes are very sensitive to changes in climatic conditions.
7. Climate Warming Considerations

The development of models of the heat and mass exchange processes for lakes is crucial for understanding the potential impacts of climate changes on the northern water resources. Considerable work has been conducted on the Laurentian Great Lakes (e.g. Schertzer and Croley, 1999). An initial investigation on potential climate impacts was conducted on Great Bear Lake (Meyer et al., 1994) using limited meteorological and lake data combined with 2 x CO$_2$ GFDL steady-state climate scenario in which it was found that under climate warming, ice cover period reduced from 85% to 65% of the year and this resulted in increased water temperature, increased convective mixing and enhanced air-water exchange. Under GEWEX-MAGS-2, lake thermal models will be combined with detailed lake time-series data and current climate scenarios to assess the potential heat and mass exchange responses of lakes such as Great Slave Lake.

8. Future Areas of Research

Under the GEWEX-MAGS-2, research will continue to improve understanding of major exchange processes and modelling.

8.1 SSM/I Ice Freeze-up and Break-up Research (Walker et al.)

Research will be conducted to improve quantification of lake ice freeze-up and break-up dates. A major effort will be to extend the ice time-series back to 1979 utilizing 37 GHz frequency data. In addition, improved spatial resolution should result from use of AMSR satellite (EOS Aqua and ADEOS-11) data, especially for smaller lakes.
8.2 AVHRR Temperature Research (Bussières et al.)

The next phase of research will include derivation of seasonal temperature trends applied to the 8900 lakes with surface areas $\geq 1 \text{ km}^2$. Advances are expected in the inclusion of weather fluctuations within the AVHRR temperature trends. The research has significant applications for model validation and initialization temperatures for the Canadian Regional Climate Model (CRCM), for the Canadian Land Surface Scheme (CLASS), for WATFLOOD and lake thermal models.

8.3 AVHRR / CERES Research for Derivation of NSSR (Leighton et al.)

Research will continue in improving the algorithms for estimation of NSSR and this will include continued inter-comparison of satellite-derived fluxes with surface measurements. Modelling results have significant applications for surface radiation modelling investigations and for validation of atmospheric models.

8.4 Lake Measurements and Modelling (Schertzer et al. and Rouse et al.)

Research will focus on development / verification of models for lake heat fluxes, evaporation, lake temperature and ice formation / decay for a range of lake sizes over the MAGS Basin. This will involve collaborative research with MAGS colleagues in remote sensing and modelling. Research will also be directed to link / couple a lake model with CLASS and to cross-evaluate temperature predictions from the CRCM. A major goal is to understand potential climate impacts on the heat and mass exchange and thermal characteristics of northern lakes.

9. References


Tracing of Water on Eurasian Continent using Stable Isotopes of Water

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1. Introduction

Water isotopes can be used for tracing water in the natural system. It has been pointed out that recycling of water has an important role for inland precipitation and land-surface interaction (reviewed in Eltahir and Bras, 1996). Water isotopes have been used for investigation of water recycling in the Amazon and Europe (Salati et al., 1979; Rozanski et al., 1982). Oxygen and hydrogen isotope ratios, which are expressed using $\delta^{18}O$ and $\delta^D$ values, decrease with distance from the ocean. This phenomenon is called the "inland effect" (Dansgaard, 1964). Recycling of water which precipitated on the surface, however, suppresses the decrease of the $\delta^{18}O$ and $\delta^D$ values.

Eastern Siberia is the farthest from a source of water (ocean) in the world, when water vapour is brought by westerly wind. Precipitation was taken at the observational site of GAME, Yakutsk, for investigation of recycling of water.

Isotopes of water in various materials such as soil and plant are also a powerful tool for tracing water in the system. During the observational period of GAME-Siberia, soil water, sap water, leaf water, lake and river water, and atmospheric water vapour were measured. These data give us valuable data for water flow or movement in the system.

2. Precipitation

Precipitation $\delta^{18}O$ on the Eurasian continent varies seasonally and spatially (Figure 1). During winter, the $\delta^{18}O$ values decreased from west to east significantly. The observational site (56°N, 12°E), which is the closest to the ocean, shows relatively constant $\delta^{18}O$ values through out a year. With greater distance inland, the difference in $\delta^{18}O$ between summer and winter becomes larger. Extremely low $\delta^{18}O$ values were observed during winter at Yakutsk (62°N, 129°E). Conversely, summer rainfall showed a very high value. This result indicates that contribution of the water recycling process for precipitation is large during summer (Sugimoto et al., in preparation).

GNIP data on North America shows a different pattern (Figure 2). Bethel (61°N, 161°W) shows higher $\delta^{18}O$ than the others, and the difference between winter and summer is the smallest. The other sites show lower $\delta^{18}O$ values than Bethel, and spatial variation from west to east exhibits a different pattern between winter and summer. In January, the $\delta^{18}O$ decreased from west to east, while it increases from west to east during summer. Freezing condition of Hudson Bay and wind direction may affect the variation of precipitation $\delta^{18}O$. 
Figure 1: Monthly mean values of precipitation $\delta^{18}$O on Eurasian continent. GNIP data and our data at Yakutsk are shown. Data for Yakutsk are 1998-2000 averaged values for May to September, and for another period obtained values during 1999-2000 winter are plotted.

Figure 2: Monthly mean values of precipitation $\delta^{18}$O in North America. GNIP data are plotted.
Figure 3 shows a decrease in the $\delta^{18}$O schematically. When no water is recycled after precipitation, water vapour $\delta^{18}$O decreased inland because heavy isotopes precipitate faster than light isotopes. This may be a similar condition to that observed at Yakutsk during winter. On the other hand, water vapour $\delta^{18}$O does not decrease if all water was recycled after precipitation (bottom in Figure 3). Summer rainfall is formed under an intermediate condition.

3. Soil and Plants

3.1 Soil Moisture and its $\delta^{18}$O

Water flow in soil can be traced using soil water $\delta^{18}$O. Soil moisture was observed (with TDR) at Spasskaya Pad experimental forest, where is an observational site of GAME-Siberia, from spring to fall in 1998, 1999, and 2000. Using those data, soil water equivalent for liquid water and ice was calculated. Inter-annual variability of soil moisture was very large, depending on the amount of summer rainfall (Sugimoto et al., submitted). During spring to early summer, surface soil water $\delta^{18}$O decreased due to infiltration of snow meltwater. Infiltration was clearly traced down to 15 cm. Surface soil water $\delta^{18}$O also showed large year to year variation. Figure 4 shows the $\delta^{18}$O of soil water observed in the beginning of August in 1997, 1998, and 1999. Variation of $\delta^{18}$O in surface soil water is very large, while the $\delta^{18}$O of soil water below 60 cm was low and rather constant. Variation in the $\delta^{18}$O in the shallow soil layer reflects the direction of water flux. During drought summer conditions (1998), soil water $\delta^{18}$O in shallow layers decreased because water with lower $\delta^{18}$O was transported upward. During wet summer conditions (1999), soil water $\delta^{18}$O increased, due to percolation of summer rainwater with high $\delta^{18}$O (Sugimoto et al., submitted).
3.2 Source of Water for Plant Transpiration

Recycling of water during summer occurs through plants, namely transpiration. Usually, it is not easy to know the depth or source of water used by plants. However, it is possible to know the source of water transpired by plants from the observation of sap water $\delta^{18}O$. Sap water has the same $\delta^{18}O$ value as source water, although leaf water enriches due to evaporation (Figure 5).

Figure 5: Change in the isotopic composition of water during transpiration. When a plant takes up water, no significant fractionation occurs. Under a steady state, $\delta$ value of water leaving from the plant is same as that of water taken up, although the leaf water is enriched due to evaporation.

Figure 4: Soil water $\delta^{18}O$ observed in the beginning of August and sap water $\delta^{18}O$ observed in August in each year from 1997 to 1999.
Water source for plant transpiration changed seasonally in eastern Siberia (Figure 6). Soon after leaf unfolding, sap water $\delta^{18}O$ decreased, because plants take up snow meltwater with low $\delta^{18}O$. During mid to late summer, source of water differs depending on the amount of summer rainfall. During wet summers such as in 1999, plants can use summer rainwater with high $\delta^{18}O$, consequently sap water $\delta^{18}O$ increases. During the drought summer conditions (1998), sap water $\delta^{18}O$ decreased, indicating large contribution of ice meltwater, which is transported from the lower part of the active layer. Soil water storage as ice in the permafrost works as a buffer, and stabilizes evapotranspiration (Sugimoto et al., submitted)

![Figure 6: Seasonal variation of sap water $\delta^{18}O$ in Larix gmelinii (Sugimoto et al., in press).](image)

4. Atmospheric Water Vapour

Water vapour in the atmosphere taken by aircraft, tethered balloon, and an observational tower was analyzed for $\delta^{18}O$ and $\delta^D$ values. A clear seasonal variation was observed: the trend of the variation in the atmospheric water vapour $\delta^{18}O$ was similar to that in precipitation. Water vapour $\delta^{18}O$ gives us direct information which indicates the source of water vapour and recycling process. Obtained data is now under the analysis process.

5. References


Flow Regimes and Flow Contribution from Sub-basins of the Mackenzie System

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1. Introduction

The Mackenzie, draining an area of 1.8 million km$^2$, discharges about 9000 m$^3$s$^{-1}$ (or 285 km$^3$y$^{-1}$) of water into the Beaufort Sea. It is the largest Canadian river that brings freshwater into the Arctic Ocean, and has influence upon the oceanic circulation as well as the seasonality of sea ice. The Mackenzie basin is considered by many GCMs as an area likely to be affected by climatic warming and in fact, the climatic record has shown an increase of over 1.5ºC for the period 1950-98. This has implications on the water balance, hence runoff generation in the basin.

Although streamflow data have been collected for decades along the main trunk and many tributaries of the Mackenzie, there is a need for the synthesis of the discharge data to establish the pattern of how much water is delivered during different times of the year from different environments. This information will be useful in the assessment of the sensitivity of runoff from various parts of the Mackenzie system to the climatic forcing. The present study addresses the problem by studying how much water is delivered at different times of the year to the Mackenzie system from its main sub-basins. The approach includes examining the mean annual discharge and the seasonal flow pattern of the major sub-basins to determine their contributions to the Mackenzie.

2. Study Area and Data

The Mackenzie basin has four physiographic regions (Figure 1). In the west, the Western Cordillera consists of a series of mountain chains and valleys or high plateaus. The east is the Canadian Shield, a rolling terrain with myriad lakes and valley-wetlands separating hillslope outcrops of Precambrian bedrock. The central zone is the Interior Plains with wetlands, lakes and vegetation that ranges from the prairie grassland in the south, through the boreal and subarctic forests, to the tundra in the north. The mouth of the Mackenzie is its delta, which consists of many distributaries and lakes.

The Mackenzie drainage is divided into several major sub-basins with different hydrological characteristics and with discharge values that are measured or that can be estimated (Figure 1). They include the Athabasca, located in the cold temperate zone of southern Mackenzie; the Peace, which is regulated at the Bennett Dam; the Great Slave that includes Lake Athabasca and the drainage from the Canadian Shield, as well as the high plains (>1500 m); the Great Bear in the shield region, dominated by the large Great Bear Lake; the low plains with many basins draining wetlands, small lakes and northern forests; the Liard, which is a large mountainous basin; and the northern mountains with a collection of smaller catchments in a subarctic, subalpine setting. The Mackenzie is gauged at the village of Arctic Red River, before the river
branches into many distributaries. For this study, discharge data from this station will be used as the total flow from the Mackenzie system.

![Map of Mackenzie Basin](image)

**Figure 1:** Physiographic subdivision and major sub-basins of the Mackenzie Basin.

Monthly and annual discharges for the Mackenzie, its major sub-basins and 23 tributaries are used in the analysis. The data spans the period 1968 to 1999, and each station has at least 7 years of record. Some glacierized basins in the mountains and several rivers on the low plains are not gauged, but empirical relationship can be obtained to estimate their annual discharges depending on the basin size. For glacierized basins with areas between 15,000 and 30,000 km², \( q = 0.0218 - (7\times10^{-7})A \), otherwise \( q = 0.1875A^{-0.29} \). For the low plain basins with an area between 15,000 and 30,000 km², the relationship \( q = 0.2178A^{-0.46} \) is used, otherwise \( q = 0.006 - (2\times10^{-7})A \), where \( q \) is specific discharge in \( \text{m}^3\text{s}^{-1}\text{km}^{-2} \) and \( A \) is basin area in \( \text{km}^2 \).

### 3. Mean Annual Flows

Data from the headwater catchments show that glacierized basins yield the highest flow on a per unit area basis, followed by the non-glacierized basins in the mountainous areas. Both the shield and the plains have low runoff. This accounts for the large runoff from the mountainous sub-basins of the North, the Liard and the Peace, and the low runoff from the shield and plains of the Great Bear and the Great Slave basins (Table 1). The Athabasca, with mountainous headwaters combined with lower flows from the high plains and the shield, has intermediate runoff values.
Table 1: Sub-basin mean annual runoff (mm) and flow contribution to the Mackenzie system.

<table>
<thead>
<tr>
<th>Basins</th>
<th>Basin Area (km²)</th>
<th>Annual Runoff (mm)</th>
<th>Percentage Contribution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern Mountains</td>
<td>112,037</td>
<td>307</td>
<td>10</td>
</tr>
<tr>
<td>Liard</td>
<td>275,000</td>
<td>279</td>
<td>27</td>
</tr>
<tr>
<td>Peace</td>
<td>293,000</td>
<td>223</td>
<td>23</td>
</tr>
<tr>
<td>Athabasca</td>
<td>307,000</td>
<td>159</td>
<td>17</td>
</tr>
<tr>
<td>Great Bear</td>
<td>145,000</td>
<td>114</td>
<td>6</td>
</tr>
<tr>
<td>Low Plains</td>
<td>138,452</td>
<td>104</td>
<td>6</td>
</tr>
<tr>
<td>Great Slave</td>
<td>404,470</td>
<td>103</td>
<td>14</td>
</tr>
<tr>
<td>Mackenzie at Arctic Red River</td>
<td>1,680,000</td>
<td>169</td>
<td>103</td>
</tr>
</tbody>
</table>

4. Regimes

Regime is the average pattern of seasonal variation in streamflow. Streamflow is influenced by water supply (e.g. snowmelt, rainfall, glacier melt), water losses (e.g. evaporation) and storage modifications (by lakes, wetlands, reservoirs and groundwater). The flow of the Mackenzie River is dominated by peak flow caused by snowmelt and amplified by river ice breakup. The annual peak is followed by declining flow in the summer and low flows in the winter. The flow pattern of the Mackenzie combines the regimes of its sub-basins, each with its seasonal pattern.

Several major regime types can be recognized. The primary seasonal flow pattern is the subarctic nival regime in which high flows are generated by snowmelt, often associated with ice breakup in the rivers. While most of the rivers in the southern basin and at low altitudes peak in May, rivers at high latitudes and high altitudes have delayed snowmelt and hence their spring peaks occur later. In glacierized basins, the ablation of glaciers is intensified in the summer and this enables the high flows to be prolonged into the summer. For some basins, autumn rainfall can give rise to a secondary peak that is lower in magnitude than the spring flow.

While wetlands have little effect in modifying the spring high flows because of their low storage capacity when frozen, the large amount of moisture available at or near the ground surface enhances evaporation in the summer. After the ground thaws, wetlands have increased ability to retain water and retard the summer flows. Large lakes are even more effective in providing large storage capacities to reduce the high flows and to extend the low flows. Thus, basins with a pro-lacustrine regime tend to have fairly even runoff during the year. When the flow is modified by reservoir operation to generate hydroelectric power, as is the case of the Peace River, the natural flow regime is strongly altered though the total annual flow volume is not seriously affected.
5. Sub-basin Flow Contributions

Different parts of the basin play a varying role in terms of percentage contribution to the Mackenzie flow (Figure 2). In the spring, much of the flow is contributed by the southern basins with early melt. In the summer, the Liard basin is the main flow contributor. The central parts of the basin yield the majority of the Mackenzie discharge during the autumn. In winter, the Athabasca and the regulated Peace River sustain much of the low flow.

The total annual flows from the main sub-basins add up remarkably close to the annual flow of Mackenzie River at Arctic Red River (within 5 percent). On a monthly basis, however, the Mackenzie River has lower flow than the combined sub-basin discharge in May, but is reversed in the summer (Figure 2). Possible reasons include:

1. errors in discharge measurement and calculation,
2. storage along the channel and in the riparian zones during the spring season, such as ponding and ice jam flooding; this water is later released to the Mackenzie flow system,
3. there may be influence and effluence along the channels which is not reflected in the river gauge measurements.

6. Summary

Within the Mackenzie Basin, there is a northward decrease in flow. This reflects the spatial trend in precipitation, which shows a decline towards the northern plains and the shield areas. The mountainous basins produce the highest annual discharges in the Mackenzie Basin and the flows exhibit the largest annual variations. On the other hand, basins in the shield and on the low plains have the lowest discharge, but they still have large seasonal variations. Large lakes even
out the seasonal flow variations and they are important in supporting the winter low flows of the Mackenzie. The highly regulated Peace River also has reduced seasonal flow variations due to hydroelectric power dam operations.

To understand the sensitivity of the Mackenzie system to climate variation and climate change, it is necessary to study not only how much, but the timing of flow contributions from its various sub-basins.
REMOTE SENSING STUDIES
Some Phenological Aspects of Vegetation over Siberia and North America as Revealed by NDVI

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1. Introduction

From the 1980s, global vegetation index data have been provided by satellite measurements from space. The NOAA satellite has a 5-channel radiometer, Advanced Very High Resolution Radiometer (AVHRR), with channel 1 in visible (580-680 nm) and channel 2 in near-infrared (735-1100 nm) spectral bands. The Normalized Difference Vegetation Index (NDVI), which is the most well known vegetation index, is computed by the equation:

\[ \text{NDVI} = \frac{\text{Ch2} - \text{Ch1}}{\text{Ch2} + \text{Ch1}} \]

where \( \text{Ch1} \) and \( \text{Ch2} \) are the reflectance measurement of AVHRR Channels 1 and 2, respectively.

This study focuses on phenological characteristics of vegetation over Siberia and North America as revealed by the NDVI, especially on the west-east contrast in each continent. Suzuki et al. (2001) examined the phenological regionality and the climate over all Siberia by using monthly NDVI data. However, the temporal resolution of the monthly data utilized in that paper was not enough to resolve the detail phenological characteristic. The present study analyzes the NDVI seasonal cycle based on weekly NDVI. The west-east phenological contrast in Siberia was investigated by Suzuki et al. (1998), while the present study surveys it over North America in addition to Siberia, and focuses on the difference between them.

2. Data and Analyses

Weekly mean values of the NDVI, averaged through 1989 to 1991 (3 years), were analyzed. The regions targeted by the analysis are bounded by 30°E and 145°E, and 40°N and 75°N for Siberia, and 140°W and 50°W, and 40°N and 75°N for North America. The second-generation weekly Global Vegetation Index (to be referred to as ‘GVI’) data, which are utilized in this study, are constructed by processing the original AVHRR measurement (e.g., Kidwell, 1990). The spatial resolution is 0.144 x 0.144 degree. This study carried out two smoothing processes, the smoothing in the horizontal domain, and then, smoothing in the time domain. For horizontal smoothing: (1) the NDVI values at a pixel (NDVI-a) were compared with the mean NDVI (NDVI-m) among the 8 pixels adjacent around NDVI-a pixel, (2) if the difference between NDVI-a and NDVI-m was greater than 0.1, NDVI-a was replaced by NDVI-m, and (3) this horizontal smoothing process was executed on all GVI pixels. For temporal smoothing, a 3-week moving maximizing process was carried out, and subsequently, weekly NDVI maps (3-year mean) from the 2nd to 52nd week were constructed for Siberia and North America.
3. Phenological Characteristics of the NDVI

As the timing parameter of the vegetation phenology, two specific weeks are picked up; green-up time defined by the week when the NDVI exceeds 0.2 for the first time in the year (W-a) and the maximum time defined by the week at the annual maximum NDVI (W-b).

3.1 Siberia

Figure 1 demonstrates the distribution of W-a and W-b. The earliest region of W-a is located in a region around the Black Sea, before the 18th week (before 30 April). From this region, W-a propagates eastward, and forms an early W-a zone around 57°N of the latitudinal line. Within this west-east zone, the W-a gradually becomes later from west to east (22nd week, 28 May - 3 June, around Baykal lake), while in the coastal area of the Sea of Japan, early W-a (19th week; 7-13 May) can be also seen. Departing northward from these early W-a zones, the W-a becomes later. The W-a in the tundra region is almost after the 28th week (9-15 July) that is 10 weeks later than around Black Sea.

![Figure 1: The distribution of the week when the NDVI exceeds 0.2 (a) and NDVI reaches annual maximum (b). The black colour stands for the pixel that has earlier week than the earliest week of the grey scale, and the white colour the pixel that has later week than the latest week of the grey scale. The W-a and W-b were not calculated in the grey areas around Caspian Sea, Aral Sea, and Mongolia because the NDVI did not exceed 0.2.](image-url)
The overall feature of the W-b distribution pattern is similar to the W-a, while some different points can be observed. The earliest region of W-b is western region (30-80°E) along the longitude of 50°N which includes Kazakh Steppe, before 21st week (21-27 June). Early weeks distributed mainly in the zone between 50 and 60°N, and in addition, there is west-east gradient, that is, the west region is generally earlier than east region. Towards the north from this early week zone, the week gradually becomes later, and the latest week in tundra area is 31st week (30 July - 5 August) that is 7 weeks later than the earliest region. W-b around Gobi desert is also generally late.

This study highlighted the west-east contrast of the phenological characteristics known from W-a and W-b in 50-60°N zone. This west-to-east propagation of the NDVI phenological events, so-called ‘green-wave’, is salient in Siberia. Figure 2 indicates the west-east variation in W-a and W-b that were averaged between 50 and 60°N. Around 40°E, the W-a is about 18th week (beginning of May), while the W-a around 112°E is 22nd - 23rd week (beginning of June), that is, the green-up of the NDVI in 112°E is one month later than that in 40°E. W-b also shows a similar zonal profile. The W-b around 40°E is about the 24th - 25th week (mid June), while the W-b around 112°E is about the 29th week (mid July), having four weeks difference.

3.2 North American Boreal Zone

Figure 3 shows the W-a and W-b distributions in North America (mid to high latitude). The earliest region of W-a is located along southeast and southwest coastal areas, before 18th week (before 30 April). The second earliest region is found around Great Lakes, 18th week (30 April - 6 May). From these regions, W-a propagates toward inland and northern areas. It is interesting that an isolated early region (mid May) can be also seen around 120-110°W and 53-60°N, the upstream area of Peace River. The latest zones, 28th week (9 - 15 July) or later than that, can be seen in the tundra areas in the west and east to Hudson Bay.
Figure 3: Similar to Figure 1, but for North America continent. The W-a and W-b were not calculated in the grey areas in arctic islands because the NDVI did not exceed 0.2.

The earliest region of W-b, 23rd or 24th week (beginning to mid June), is in southern regions except for south of the Great Lakes. From this early week zone toward the north, the week gradually becomes later, and the latest week in tundra and northeast coastal areas is the 31st week (30 July - 5 August) that is 7 weeks later than in the earliest region. W-b around the Rocky Mountains is relatively early compared with tundra area.

Figure 4 indicates the mean W-a and W-b in the zone between 50 and 60°N of North America. West-east contrast is not clear, while coast-inland contrast is rather outstanding in W-a, namely the early W-a is found around 120°W and 88°W in this figure. As for W-b, west-east variation is not apparent as demonstrated in Figure 4, that is, W-b is in mid to late July for all longitude.
Figure 4: Similar to Figure 2, but for North America continent.

4. Discussion and Summary

The phenological characteristic of Siberia and North America boreal zone (especially in 50-60ºN zone) was investigated by using 3-year mean remotely-sensed NDVI data on weekly basis. The NDVI green-wave propagating west to east was found in 50-60ºN zone of Siberia. This fact delineates that there is a phenological 4-5 week time lag between western and eastern region of Siberia.

However, such west-to-east green-wave does not exist in North America. Instead of that, coast-inland contrast is obvious. Although we should take in account the considerable difference of west-east extent of Siberian and North America boreal areas, there should be other geographical backgrounds which cause the difference between the phenology in Siberia and North America.

We consider that the main reason for no green-wave in North America is the difference in topographical and the climatological backgrounds between Siberia and North America. The Rocky Mountains are located in the west coast of North America. Although Siberia has the Ural Mountains in the East, the height of the Ural Mountains is much lower than that of the Rocky Mountains. Also, the large water bodies of the Great Lakes and Hudson Bay, which break into the terrestrial area of continent, could moderate the climate. The isolated area that shows irregularly early W-a over the upstream area of Peace river corresponds to the ‘small leaf mixed wood’ area according to the USGS’s land cover data base (Loveland et al., 2000). This fact implies the importance of the vegetation type for the phenology in addition to topography and climate.

This study pointed out the phenological difference between Siberia and North America by using the global NDVI data. To elucidate the phenological difference between them and its reason, further surveys and discussion should be required to reveal the difference.
5. Acknowledgements

All pictures were constructed with the help of the GMT System (Wessel and Smith, 1991).

6. References


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Abstract

The force restore method was originally developed for estimating diurnal fluctuations in ground surface temperature. Because of its relatively simple parameterization, it is commonly applied in meteorological and other models for this purpose. Its application to the calculation of deeper soil temperatures, frozen soils and soils under snow cover has heretofore not been possible. This study demonstrates an extension of the force-restore method that permits accurate estimates of seasonal variation in deep soil temperature; frozen soil depth and ground surface temperature under a snow cover. The extended method is tested using measurements collected in a cold continental region. The modified formula can also be applied to determine the lower boundary condition for calculations of diurnal variations in soil temperature.

1. Introduction

The force restore method (FRM) is an alternative approach, developed to estimate the ground surface temperature. Hu and Isram (1995) showed that the FRM could provide accurate estimates of both ground surface and upper soil temperatures by minimizing the error between the analytical solution from the force-restore method and from that of the heat conduction equation under diurnal forcing. Hirota et al., (1995) demonstrated and tested with field measurements in Japan, an extension of the FRM to estimate the seasonal variation in daily mean soil temperature of shallow (upper) soil layers. However, these versions of the FRM did not consider estimating deep soil temperature, nor have they been fully tested under snow cover and for frozen soils. It is doubtful that existing FRM formulations can accurately represent the ground surface temperature under a snow cover as they assume a strong diurnal ‘forcing’ at the surface. Insulation of soil from air by snow cover will strongly dampen diurnal temperature fluctuations at the ground surface and may violate the force-restore assumption. The prospects of successfully applying an unmodified FRM approach to cold, snow covered regions are therefore questionable.

This study proposes a new and simple method for estimating deep soil temperature using a modification of the FRM. The FRM can predict diurnal variations in ground surface temperature if appropriate boundary conditions can be specified. Here, extended methods for seasonal variations in mean daily deep soil temperature are shown. The new method also can be applied to determine lower boundary condition of soil temperature to estimate ground surface
temperature or vertical soil temperature profile. It can be applied for not only mean daily value but also diurnal variations. This application of the FRM is also effective for estimating daily boundary conditions and estimating soil temperature profile for frozen and unfrozen soil under snow. The result of the extended FRM is compared to measurements of soil temperature regimes over a winter near Saskatoon, Saskatchewan, Canada.

2. Extension of the Force Restore Method

2.1 Review of Method

The sinusoidal soil temperature changes is expressed by

\[ T(z,t) = \bar{T} + \Delta T_0 e^{-\frac{2\pi^2}{\omega^2 D_a}} \sin \left( \frac{\omega t - z}{D_a} \right) \]  

(1)

where \( T(z,t) \) is the soil temperature over some vertical coordinate, \( z \), and time \( t \), \( \bar{T} \) is the mean ground surface temperature (daily or annual), \( \Delta T_0 \) is the daily or annual temperature amplitude at the surface, and \( \omega \) is the frequency of oscillation equal to \( 2\pi / \tau \). \( D_a = (2\pi / \omega \alpha)^{0.5} \), \( D_a \) is the damping depth (m) of surface temperature fluctuations, \( \tau \) is the period of temperature fluctuation calculation (day or year), \( \alpha \) is the soil thermal diffusivity, found as \( \alpha = \lambda / c \), \( \lambda \) is the soil thermal conductivity, \( c \) is the volumetric heat capacity.

The vertical conductive heat flux in a soil, \( G \), at depth \( z \) and time \( t \) is given by

\[ G(z,t) = -\lambda \frac{\partial T(z,t)}{\partial z} \]  

(2)

Combining Eqs. 2 and 1 provides

\[ G(z,t) = \left( \frac{\omega \alpha \lambda}{2} \right)^{0.5} \left( \frac{\omega}{\lambda} \frac{\partial T(z,t)}{\partial t} + T(z,t) - \bar{T} \right) \]  

(3)

which is a differential form of the soil heat flux with respect to time (t). Eq. 3 provides the soil heat flux differentiated with respect to distance and time (Bhumralkar, 1975).

2.2 Application to the Soil Surface Layer

Considering a soil surface layer of thickness below the ground surface, as shown in Figure 1, then the rate of temperature change over time for this layer is given by

\[ c \frac{\partial T_s(\delta,t)}{\partial t} = -\left( \frac{G(\delta,t) - G(0,t)}{\delta} \right) \]  

(4)
T_g, the ground surface temperature is defined as

$$T_g(\delta,t) = \frac{1}{\delta} \int_0^\delta T(z,t) \, dz$$  \hspace{2cm} (5)

Assuming that $T(\delta,t) \approx T_g(\delta,t)$, and combining Eqs. 5 and 4, we obtain

$$C_1 \frac{\partial T_g(\delta,t)}{\partial t} = \frac{2}{cD_a} G(0,t) - \frac{2\pi}{\tau} (T_g(\delta,t) - \overline{T})$$  \hspace{2cm} (6)

Here, $C_1$ is the function of $\delta$ and $D_a$. Hu and Isram (1995) developed $C_1$ function to minimize the difference between the analytical solution of the FRM and the full heat conduction equation (HCE) in response to a single periodic forcing. Their polynomial approximation is

$$C_1 = 1 + 0.943(\delta / D_a) + 0.223(\delta / D_a)^2 + 1.68 \times 10^{-2}(\delta / D_a)^3 - 5.27 \times 10^{-3}(\delta / D_a)^4$$  \hspace{2cm} (7)

For case of $\delta \to 0$ in Eq. 7,

$$\frac{\partial T(0,t)}{\partial t} = \frac{2}{cD_a} G(0,t) - \frac{2\pi}{\tau} (T(0,t) - \overline{T})$$  \hspace{2cm} (8)

This, in the same form as the original, is a FRM of ground surface temperature for which $T(0,t) = T_g(0,t)$.

![Figure 1: Schematic diagram of the Force-Restore for a soil surface.](image-url)
2.3 Extension to Mean Daily Soil Temperature

The FRM can be applied from Eq. 8 to estimate variations in ground surface temperature, however its extension to calculation of soil temperature has been restricted for several reasons. It has been maintained that determining the daily mean ground surface temperature is problematic in that a value of $\bar{T}$ is required before solving for the diurnal variations of soil temperature using the FRM (e.g., Mihailovic et al., 1999). In addition, $\bar{T}$ is required at depth to provide a lower boundary condition for diurnal calculations. The value of $\bar{T}$ may also have to respond to changing surface thickness, $\delta$ from day to day.

Application to annual variations may be less of a problem as field measurements show that the mean annual soil temperature, $\bar{T}_{ym}$, is relatively invariant with depth. Therefore $\bar{T}_{ym}$ can be treated as a constant for a location, and its value need not be changed with changing $\delta$. $\bar{T}_{ym}$ at a location can be estimated using well-tested, simple empirical equations from the mean annual air temperature, permitting a relatively easy parameterization of the FRM for calculations of annual variation in mean daily soil temperature (Hirota et al., 1995).

2.4 Extension to Mean Daily Soil Temperatures at Depth

When applying Eq. 4 to an internal soil layer to estimate daily mean soil temperature as shown in Figure 2, then the time rate of temperature change for this layer is given by

$$c \frac{\partial \bar{T}(z,t)}{\partial t} = - \left( \frac{G_n - G_{n-1}}{\delta} \right)$$

(9)

Combining Eqs. 3 and 9 provide

$$C1 \frac{\partial \bar{T}(z,t)}{\partial t} = - \frac{2}{cD_a} G_{n-1} - \frac{2\pi}{\tau} (\bar{T}(z,t) - \bar{T}_{ym})$$

(10)

Here, $\bar{T}(z,t)$ is the daily mean soil temperature, $\tau_y$ is the annual period (365 days) and $G_{n-1}$ is the daily mean soil heat flux between an upper and internal soil layer; expressed as follows

$$G_{n-1} = -\lambda \frac{\partial \bar{T}(z,t)}{\partial z_1}$$

(11)

Combining Eqs. 10 and 11 obtains

$$C_1(\delta) \frac{\partial \bar{T}(z,t)}{\partial t} = - \frac{2\lambda}{cD_a} \frac{\partial \bar{T}(z,t)}{\partial z_1} - \frac{2\pi}{\tau} (\bar{T}(z,t) - \bar{T}_{ym})$$

(12)

where $z_1$ is the distance from upper soil depth to internal soil depth.
Note that when solving the heat conduction equation (HCE), \( \frac{\partial T}{\partial t} = \frac{\lambda}{c} \cdot \frac{\partial^2 T}{\partial z^2} \), for annual variations in soil temperature, it is necessary to set a lower boundary condition at several to several tens of metres depth. At the lower boundary, soil temperature is constant or the soil heat flux is zero. However, the extended FRM using Eq. 12 does not need to make such assumptions about deep soil conditions, permitting flexible lower boundary conditions that can be provided to a calculation such as the HCE.

3. Application of the Extended FRM

3.1 Comparison to the Heat Conduction Equation Given Set Boundary Condition

The extended FRM for deep soil temperature calculation was compared to an analytical solution to the heat conduction equation \( \{\text{HCE (Eq. 1)}\} \). This comparison used a sinusoidal soil surface temperature forcing from Eq. 1 when \( z=0 \). A comparison of results from the analytical solution to the HCE and the extended FRM (Eq. 12) under given boundary conditions (\( \bar{T} = 10^\circ\text{C}, \Delta T_0 = 30^\circ\text{C}, D_a = 2 \text{ m} \)) is shown in Figure 3. The extended FRM result coincides extremely closely to the solution of the HCE.
Soil temperatures below depths of approximately 0.3 to 0.5 m can be treated as daily constants for calculating diurnal variations of soil temperature by the HCE. Diurnal changes in soil temperature below 0.3-0.5 m depth need not be considered, permitting this layer to form a lower boundary condition for the HCE. Eq. 12 provides a method to calculate this lower boundary condition for diurnal soil temperature calculations (Figure 4). Application of Eq. 12 below depths of 0.3-0.5 m does not require initial temperature values or soil thermometric parameters of deep soil layers. A comparison of ground surface and soil temperature calculations between calculations using a 20 layer HCE set boundary conditions at 10 m, and a 6 layer HCE using boundary conditions at 0.5 m from Eq. 12 is shown in Figure 5. Simulation was carried out one-year period. The differences are within 0.07 °C and 0.8 °C for ground surface and 0.5 m deep soil respectively.

Figure 3: Comparison of soil temperature between the analytical solution of the heat conduction equation and the extended Force-Restore Method ($z_I = 0.05$ m).

3.2 Combined Method for Estimating Diurnal Variation in Soil Temperature

Figure 4: Concept of combination of heat conduction equation and the proposed method.
4. Estimating Soil Temperature under Snow Covers

Snow is an excellent thermal insulator and under non-melting snow covers, the soil surface temperature is controlled more by soil temperatures than by atmospheric conditions, because the thermal conductivity of soils is generally greater than that for snow. However, the FRM presumes significant heat exchange between soil and atmosphere by assuming a periodic temperature forcing at the surface. This assumption clearly is not valid for soils under snow covers.

Accepting these difficulties, application of the following models to soil under snow is attempted:
1. HCE model with the boundary condition given as the temperature observed at 0.8 m depth,
2. the original FRM (Eq. 6) for 0.025 m depth, and
3. the extended FRM coupled to HCE to determine the lower boundary condition 0.4 m depth.

All models were run to calculate mean daily soil temperature by using mean daily meteorological values. As mean daily winter energy inputs at the surface are comparatively small, it was assumed for the purposes of these calculations that the effect of mean daily net radiation, and latent heat flux at a ground surface under snow were negligible.

4.1 Results

Measurements used in this study were collected from Kernen Farm, located a few kilometres east of the city of Saskatoon (52ºN, 107ºW) in the central southern half of the Province of Saskatchewan, Canada. Figure 6 shows soil temperatures observed and calculated by three methods. The first method (estimated by full HCE) and third method (extended FRM coupled to reduced HCE) agree well with observed values. Root mean square errors are 1.9ºC by the first method, and 1.6ºC by the third. The second method (using original FRM) did not agree well with observations during the snow-covered period, with a root mean square error of 4.6ºC. These results suggest that the mean daily soil temperature under snow cover in mid-winter can be estimated from mean daily air temperature, snow density and snow depth, without considering net radiation and latent heat.
During the snowmelt season, from the end of February to the beginning of March, all modelled values were underestimated compared to observations. The underestimation is likely due to the model implementation not considering the additional energetics of infiltrating meltwater into frozen soils (Pomeroy et al., 1998). Any model of soil temperature will need to consider the effect of meltwater infiltration to frozen soils to accurately estimate soil temperatures during the melt period.

The boundary condition of the third method is at 0.4 m depth. To estimate soil temperatures at depths below this, Eq. 12 can be used to extrapolate downward. Figure 7 shows the resulting comparison of observed soil temperatures at 0.8 m and 1.6 m depth and estimated values by using the 0.4 m estimated value by the third method and extrapolating using Eq. 12. Estimated soil temperatures matched observations reasonably well. Therefore this method can also be used to estimate deep soil temperatures in frozen conditions.
5. Conclusions

A simple formula to estimate seasonal variations in deep soil temperatures was developed using the force-restore approach. This formula can be used to accurately calculate the daily mean soil temperature at depth without considering deep soil thermometric conditions. Its advantages are a substantial savings in computational capacity and easier parameterization than full, multi-layered heat capacity equation calculations. The force-restore approach was extended from calculation of ground surface temperature to the calculation of deep soil temperature and frozen soil depth. This was demonstrated in a cold continental climate by using observations from Saskatchewan, Canada. To apply the method accurately to a wide variety of surface and climate conditions it needs to be coupled to a surface energy balance model and to consider the effect of meltwater infiltration to frozen soils on soil temperature.

6. References


Hydrological Modelling of Large Northern Rivers

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1. Introduction

The Arctic Ocean and its marginal seas are key areas for understanding the global climate system and its change through time. About 80% of the total terrestrial runoff to the Arctic Ocean is supplied by the Ob, Yenisei, Lena and Mackenzie rivers (Aagaard and Carmack, 1989). According to the GAME-Siberia project data set measured in 1986, 1987 and 1988, the freezing occurs in late October near the outlet, early in November midstream and in late November upstream. On the other hand, thawing begins from upstream in late April and then extends into midstream and downstream. Downstream, the river thawing occurs after June (Ma and Fukushima, 2001). Ma et al. (2000) made a hydrological simulation using a combined model to understand the hydrological processes of the Lena River basin. Kite et al. 1994 and 1999 had hydrological simulations for the Mackenzie River basin. However, river freezing processes were not considered in those models. Fukushima and Ma (1999) suggested that river ice processes should be considered for the hydrological modelling of northern rivers.

There are two basic methods of predicting ice phenomena in rivers (Michel et al. 1986). The first uses past records (e.g., Laszloffy, 1948, Liser 1959, Deslauriers 1966, Gerard and Karpuk, 1979); the second uses mathematical or empirical relations based on physical principles (e.g., Shulyakovskyii, 1963, Williams, 1965, Michel, 1971, Ashton, 1973, Ashton, 1979, Greene, 1981 and Beltaos, 1984a). A more rigorous solution of river ice processes should account for the influence of snow cover, radiation, evaporation and convection in the air, and advection in the water, however the required climatic data are seldom available. For most slow-flowing rivers, though, the condition of water and ice system can be regarded as a state of relative rest. Therefore, thermal effects dominate the ice growth and decay processes only. In this case, river ice thickness can be predicted using local hydrological and meteorological data (e.g. air temperature, radiation and water temperature).

Ma et al. (2001) conducted a river ice-cover process study, in which a simple, accumulated degree-days method was used to estimate ice-cover growth and decay events using daily routine meteorological data. Their results showed that the date of river ice breakup could be simulated well at 43 sections over the Lena basin. Here, a one-dimensional framework which considers the impact of river freezing processes on the river flood formation is summarized to explain how hydrological processes in cold regions are simulated. Application of the model is carried out in the Lena River basin and comparisons between the modelling result by Ma et al. (2000) and that of the study of Ma and Fukushima (2001) are shown.
2. Model Structure

The model is composed of four components: one-dimensional soil-vegetation-atmosphere transfer (SVAT) model, runoff formation model, river ice model and river routing model. The SVAT model, runoff formation model and river routing without a freezing factor were described by Ma et al. (2000) and the river ice model was described in detail by Ma et al. (2001). Figure 1 shows the model structure. The input of the model system is daily meteorological data and the items of output contain heat fluxes on land surface, active layer depth, hydrograph at watershed outlet and so on. A brief summary of each model component is presented below in sections 2.1 - 2.4.

![Model Structure Diagram]

Figure 1: Flowchart of the modelling system (after Ma and Fukushima, 2001)
2.1 SVAT Model

The SVAT model is a simple biosphere model, in which the land surface is described by a big-leaf and a soil layer. Using daily meteorological data, the model provides estimates of latent and sensible heat fluxes between the land surface and the atmosphere, and thermal regimes in the snow-cover and soil layer.

2.2 Runoff Model for each Grid

A conceptual model (Fukushima, 1988, p.89-91) is used to determine the formation of runoff for each 1\degree x 1\degree grid box. There is a reservoir system representing each of the four runoff components, which are saturated land surface runoff, infiltration runoff from the topsoil zone, base runoff and direct runoff from the water surface. In order to consider the permafrost condition, a parameter-related mean effective soil depth was designed as a function of the active layer depth (Ma et al., 2000). The estimated runoff for each grid will be used as one of input data by the river routing model.

2.3 River Ice Model

A simple method using accumulated degree-days is used to determine river ice growth and decay processes. The method is often used in practice to study river ice (e.g., Greene, 1981; Michel et al., 1986 and Shen et al., 1991). Here, we assume that the river ice is growing when the air temperature is below the freezing point and the river ice decay process will be started when the growth process breaks off. The model was applied to the Lena River basin (Ma, et al., 2001) and checked by a ten-day data set of ice thickness during freeze-up period, including 51 sections of the rivers over the basin (Figure 2). The result showed that for about 60 per cent of all river sections, the estimated breakup dates are consistent with observed ones.

2.4 River Routing Model

A constant velocity of 0.4 m s\(^{-1}\) for the river network of the Lena River basin with a 1\degree x 1\degree grid (Ma et al., 2000) was assumed. Actually, water flows in the channel system is very complex, especially during river ice breakup season. Here, it is designed that most of the snowmelt is stored temporarily in the river while the river fully freezes. Once the river ice breaks, part of the stored runoff over a depth is moved along the river with a high velocity (e.g. Doyle and Andres, 1978; Parkinson, 1982; Beltaos, 1984b; Gerard et al., 1984; Prowse, 1994), considering the amount of hydraulic pressure from snowmelt stored in the channel. Here, a relationship of storage and runoff on the channel is described by a non-linear reservoir (see Ma and Fukushima (2001) in detail).

Through those procedures mentioned above, simulated hydrographs at a watershed outlet can be derived.
3. Model Application

The model is applied to the Lena River basin (Ma and Fukushima, 2001) shown in Figure 2. The duration of application is set to be the same period as that of Ma et al. (2000), which is from 1 October 1986 to 30 September 1987. For this study, two data sets are required: (1) runoff data set, and (2) river ice breakup date data set for each grid over the basin. Here, a runoff data set of 1º x 1º calculated by Ma et al. (2000) linking SVAT model and runoff formation model is used. Spatial distribution of river ice-cover broken date over the basin with a 1º x 1º grid size (Figure 3) is estimated using the river ice model and 1º x 1º gridded meteorological data derived by Ma et al. (2000) over the Lena River basin. Two data sets mentioned above are input into the river routing model.

Figure 2: Map of the Lena River basin showing the locations of hydrological stations and ice depth measured river sections. (after Ma and Fukushima, 2001)
Figure 3: Estimated spatial distribution of river breakup date over the Lena River basin in 1987. (after Ma and Fukushima, 2001)

Figure 4 shows the daily hydrographs calculated by Ma et al. (2000) at six hydrological gauges. Stations No. 3, No. 6 and No. 8 are the main river sections, and are located upstream, midstream and downstream, respectively. Other sections (No.41, No.42 and No.57) are located at the three branch rivers. Figure 5 shows the result of Ma and Fukushima (2001) considering river-freezing processes. Comparing Figure 4 with Figure 5, we see that a lot of the aspects have been improved: (1) the timing of flood rise is close to the observed one for all stations, (2) the estimated flood peak seems to be in good agreement with that observed except No.3, and (3) the calculated hydrograph is more reasonable for all stations. An evaluation of root mean square error (RMSE) for each simulation is shown in Table 1. It is clear that the accuracy of modelling of this study is improved compared to the result of Ma et al. (2000).

The hydrograph of No. 3 in Figure 5 shows that the calculated flood peak is larger than that observed. The watershed area of No. 3 is the smallest in of the six, and occupies 5% of the basin. The grid resolution of forcing data that were derived from 40 meteorological stations is 1º x 1º. This may be the main reason causing some of the errors, especially in small areas. However, the result of No. 57 in Figure 4 seems good due to the river being controlled by a dam upstream. The dam effect was not considered in this study, so part of the snowmelt was not stored in the reservoir and released later. Moreover, river flow during the ice breakup period is also influenced by some man-made processes such as a use of bomb for reducing floods, reducing flooding. The preliminary results show that the model system has an effective use in simulating hydrological processes for cold regional river basins.
Figure 4: Comparison of daily runoff between the observed and calculated of Ma et al., 2000 at the six hydrological stations in the Lena River basin from October 1986 to September 1987. The number indicates each watershed area and the percentage inside the brackets indicates the fractional area of each watershed to the whole area of the Lena River basin. (after Ma and Fukushima, 2001)

Figure 5: Comparison of daily runoff between the observed and calculated of Ma and Fukushima, 2001 at the six hydrological stations in the Lena River basin from October 1986 to September 1987. The number indicates each watershed area and the percentage inside the brackets indicates the fractional area of each watershed to the whole area of the Lena River basin. (after Ma and Fukushima, 2001)
Table 1: Values of root mean square error for the Lena River hydrological simulation

<table>
<thead>
<tr>
<th>Station</th>
<th>No.3</th>
<th>No.6</th>
<th>No.8</th>
<th>No.41</th>
<th>No.42</th>
<th>No.57</th>
</tr>
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<tbody>
<tr>
<td>RMSE1</td>
<td>0.9815</td>
<td>0.3316</td>
<td>0.5314</td>
<td>0.6595</td>
<td>0.4743</td>
<td>0.4266</td>
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<tr>
<td>RMSE2</td>
<td>0.9282</td>
<td>0.3956</td>
<td>0.2577</td>
<td>0.4761</td>
<td>0.3929</td>
<td>0.3842</td>
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<tr>
<td>ΔRMSE</td>
<td>5.4</td>
<td>24.8</td>
<td>22.5</td>
<td>28.2</td>
<td>17.2</td>
<td>9.9</td>
</tr>
</tbody>
</table>

RMSE1 = RMSE of the result of Ma et al., 2000
RMSE2 = RMSE of the result of Ma and Fukushima, 2001
ΔR = RMSE1 - RMSE2

4. Conclusions

A hydrological model for cold regions was constructed by considering river-freezing process in the river flow formation based on the Ma et al. (2000) model. A simple river ice model was used to estimate ice growth and decay processes. The output provided by the model is river ice breakup date, which was used by a river routing model to calculate river flow formation process. An application of the model system was made in the Lena River basin (Ma and Fukushima, 2001) using the estimated runoff (Ma et al., 2000) and estimated river ice breakup date over the basin. The results show that the modeled hydrographs have been improved in comparison to that of Ma et al. (2000) at six stations from 1 October 1986 to 30 September 1987. The timing of the calculated floods for all stations are closer to that observed. For most of the stations, the estimated hydrographs are reasonable, not only in terms of the flood peak but also in terms of the sharpness of the hydrographs. Although the simulation duration is limited to one year, it can be expected that the model system will be an effective tool in making hydrological simulation for northern rivers in the future.

5. References


Greene, G.M., 1981: Simulation of ice-cover growth and decay in one dimension on the upper St. Lawrence River. NOAA Technical Memorandum ERL CLERL-36, PB82-114208; 87.


Issues in Automated Watershed Parameterization

Lawrence W. Martz

Department of Geography, University of Saskatchewan, Saskatoon, Saskatchewan, Canada

Abstract

This paper examines some of the issues in automated watershed parameterization for hydrological modelling that have been addressed under the Mackenzie GEWEX Study (MAGS). The paper first reviews the scope and approach of the TOPAZ software designed for the automated segmentation and parameterization of watersheds from grid-type digital elevation models. This software was used in the studies under discussion because it is robust and comprehensive in its analysis and because it is representative of the best technology available for these purposes. The paper then examines three major issues in automated parameterization: (1) the impact of grid cell resolution on segmentation and parameterization results, (2) the impact of the scale of segmentation – the number of subwatersheds delineated – on hydrologic model performance and (3) limitations in upscaling grid cell values to subwatersheds or hydrologic model response units.

The slides used for this presentation follow:
TOPAZ

- Topopgraphic PArameteriZation: software tool for the automated segmentation and parameterization of watersheds from raster digital elevation models
- Typical of available methods
- Some issues:
  - Cell resolution
  - Number of subareas
  - Aggregation of cells (parameterization of subareas)
TOPAZ Processing Schematic

Flow vectors by the D-8 Method
Upslope Contributing Area

Channel Network Delineation

CSA (critical source area):
- threshold upslope contributing area to initiate a source channel

MSCL (minimum source channel length):
- shortest source channel permitted in network
Other raster processing:  
*distance to channel*

Flow routing sequence

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<th>NEXT NODE</th>
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Sub-basin parameterization:
*aggregation from cells to planes*

Sub-basin Length Calculation

\[
L = \frac{\sum D_i}{n} \quad \text{or} \quad L = \frac{\sum D_i}{m} \quad \text{or} \quad L = \frac{\sum D_i C_k}{\sum D_i}
\]

- \( L \): plane length
- \( D \): distance to channel
- \( C \): catchment area
- \( n \): number of subwatershed cells
- \( m \): number of subwatershed divide cells
Walnut Gulch Watershed – 183 subwatersheds
Garbrecht, Goodrich and Martz 1999

Variable-scale automated segmentation:

number of ASAs (sub-basins)
Interface between TOPAZ and the SLURP hydrological model

- DEM
- TOPAZ
- Physiographic Data
- SLURP Command & Other Files
- SLURPAZ
- Channel data
- Land Cover
- Station Locations

Variation in water budget terms with # of ASAs (subwatersheds)

- Precipitation
- Evapotranspiration
- Storage
- Comp Q
Average within-ASA variance by # of subwatersheds (ASAs)

- **Elevation**
  - Average elevation variance of hydrologic zones (10,531)

- **Distance to Channel**
  - Average distance to channel variance of hydrologic zones (119,075)

Sub-basin parameterization: *aggregation from cells to GRUs*

- Detailed flow information from maps or analysis of high resolution DEM
- Aggregate flow information
- Approaches
  - elevation averaging
  - weighted vector averaging
  - "expert system" (WATFLOOD)
Mackenzie Basin

Cell resolution impacts

- Mackenzie and Missouri Basins
- GTOPO30 base data
- Aggregate to 2, 4, 8, 16, 32, 64 km resolution
- Apply TOPAZ to segment and parameterize
### Mackenzie Basin Area Properties

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<tr>
<th>Mackenzie resolution (km)</th>
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<th>Number of sub-basins</th>
<th>Average area of sub-basins (km²)</th>
<th>Mean elev. (m)</th>
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Thank you
Impact of Frozen Ground and Snow on the Water and Energy Cycles in Eurasia

Kumiko Takata

Frontier Research System for Global Change, Yokohama, Japan

1. Introduction

The impacts of land surface processes in cold regions on large-scale energy and water cycles have attracted attention in the recent 10 years. Two sub-projects of PILPS (Project for Intercomparison of Land surface Process Scheme) are targeting cold regions: PILPS 2(d) is conducted at Valdai, Russia by a one-dimensional experiment (Schlosser et al., 2000), and PILPS 2(e) is conducted in Torne/Kalix river basins by a two-dimensional experiment (Bowling et al., 2001). In PILPS 2(d), the sensitivity of snowpacks to downward longwave radiation is examined (Slater et al., 2000). The importance of snow sublimation is pointed out in the results of PILPS 2(e) (Bowling et al., 2001) and further examination will be done. In this paper, the global-scale impact of frozen ground is reviewed, the sensitivity on snow albedo is examined, and direction for future work is discussed.

2. Global Impacts of Frozen Ground

The response of frozen ground climate has been investigated in various studies (e.g., Anisimov and Nelson, 1997; Romanovsky and Osterkamp, 1995), but an interactive effect of soil freezing and the atmosphere has not been examined before Viterbo et al. (1999). They pointed out that introducing soil freezing into the European Centre of Medium-Range Weather Forecasts model eliminates to a large extent the systematic 2 m temperature bias in winter due to the thermal inertia of freezing and thawing. They also found a positive feedback between the boundary layer and the land surface; that is, a stable boundary layer is induced by surface cooling, which reduces the downward heat flux resulting in further surface cooling. This positive feedback is relaxed by introducing the soil freezing.

Takata and Kimoto (2000) examined the impact of soil freezing on continental-scale seasonal change in water and energy cycles using an atmospheric general circulation model (AGCM). They compared the results with and without soil freezing, i.e., assumed super-cooled soil moisture when below the freezing point, and neglected the latent heat of fusion and the impermeability of frozen soil. It was shown that the inclusion of soil freezing induces higher runoff in spring and lower surface soil moisture in summer in the mid to high latitudes over land. This leads to higher surface temperature, lower evaporation, and lower precipitation implying the in situ effects of soil freezing. On the other hand, for the remote effects, the eastward water vapour flux and precipitation in Southeast Asia is enhanced in summer. It is accompanied by the similar changes in the outgoing longwave radiation and the upper atmospheric temperature to those in the strong monsoon years (e.g., Li and Yanai, 1996).
3. Large-scale Sensitivity on Snow Albedo

3.1 Model and Experiment

It is widely recognized that snow albedo has an appreciable effect on the atmosphere. A sophisticated parameterization of snow albedo has been proposed (e.g., Wiscombe, and Warren, 1980) and its sensitivity on the land surface and the atmosphere is examined using an AGCM.

The land surface model used is the Minimal Advanced Treatments of Surface Interaction and RunOff (MATSIRO). Figure 1 shows a schematic diagram of MATSIRO, and features of the model are described below.

Figure 1: Schematic diagram of MATSIRO. \( H \) is the sensible heat flux; \( E \) is the evapotranspiration; \( T_c, W_c \), and \( T_s \) are the canopy temperature, the canopy water, and the ground surface temperature respectively where superscript \((sn)\) denotes those in the snow-covered portion: \( T_{sn}, T_g \), and \( W_g \), and \( F_g \) are the snow temperature, the soil temperature, the soil moisture, and the frozen soil moisture respectively with the number of soil layer (1-5); and \( Ro \) is the runoff.
The coefficients for eddy transport and radiative transfer are parameterized from a multi-layer canopy model (Watanabe, 1993; Watanabe and Ohtani, 1995). The stomatal resistance is estimated on the basis of photosynthetic reactions (Sell"ers et al., 1996). Energy balances are solved at canopy and ground surfaces for snow-covered and snow-free fractions in a grid cell, then the weighted mean fluxes are given to the atmosphere.

The sub-grid snow fraction is diagnosed from the snow amount. The snow amount is calculated from the balance equation of snowfall, sublimation and snowmelt taking into account the refreezing of snowmelt and the freezing of rainfall. Snow has variable number of layers (1 to 3) in accordance with its amount, and the snow temperature is calculated from a heat conduction equation. The snow albedo is prognosticated, considering the reduction effects of 'aging' that depends on the time passage since the last snowfall and the snow temperature (Wiscombe and Warren, 1980).

Interception and evaporation of canopy water are incorporated. The simplified TOPMODEL (Beven and Kirkby, 1979) is implemented for runoff. Soil has 5 layers for the temperature and the wetness, calculated from the heat conduction and the hydraulic potential gradient, respectively. The soil moisture is also changed by the soil surface evaporation and the root uptake. The latent heat of freezing and thawing and the reduction in hydraulic conductivity due to soil freezing are considered. The detailed description is given in Takata and Emori (2001).

CCSR/NIES (Center for Climate System Research, University of Tokyo/National Institute for Environmental Studies) AGCM (Numaguti et al., 1997) is used at the horizontal resolution of T42 with 20 vertical layers. The experiment with a prognostic snow albedo (C run) is performed using the prescribed monthly sea surface temperature from 1982 to 1988. The experiment with a fixed snow albedo at new snow value (A run) is also conducted. The differences in the mean seasonal cycle between the two experiments are examined.

4. Results

The simulated snow covered areas in C run agree with those derived from the satellite data of NOAA/AVHRR (Figure 2). The annual mean difference (A-C) in snow albedo is high near the south edges of snow-covered regions. However, those of net shortwave radiation which affects the atmosphere is high in the regions where the snow-covered period is long, e.g., Tibetan Plateau and Rocky Mountains, and where the effects of canopy on snow albedo is low (namely, leaf area index (LAI) is low), e.g., tundra and grassland regions (Figure 3).
Figure 2: (a) Satellite observed snow cover fraction (NOAA/AVHRR) in the middle of April (top), May (second), June (third), September (fourth), October (fifth) and November (bottom). (b) as in (a) but for simulated snow water equivalent. Unit: kg m$^{-2}$. 
The time-latitude distributions of the zonal mean difference over land are shown in Figure 4 to examine the seasonal cycle of the difference between the two experiments. The difference (A-C) in snow albedo is higher in autumn (Figure 4a) than in spring. However, A-C of upward shortwave radiation is higher in spring (Figure 4c), leading to the lower spring surface air temperature in A run than in C run (Figure 4d). It is found that the positive A-C of upward shortwave radiation agrees much more with the positive A-C of snow fraction (Figure 4b) than the positive A-C of snow albedo (Figure 4a). It is thus presumed that the snow fraction has an appreciable impact on the atmosphere. The low spring surface air temperature in A run leads to a delay in snowmelt and hence a delay in the corresponding runoff (Figure 4e).

The negative A-C of surface air temperature spreads over most parts of Eurasian and North American continents in April. It continues to appear in July in Northwest Siberia, Tibetan Plateau, Eastern Europe, and the northeast of Northern America (Figure 5). The impact on the atmospheric temperature is not limited near the surface, but extended in the lower troposphere by summer (Figure 6).
Figure 4: Time latitude cross-sections of zonal mean over land. Abscissa denote the day from 1 January to 31 December. (a) Difference (A-C) in snow albedo. (b) A-C in snow fraction. (c) A-C in upward shortwave radiation. Unit: W/m². (d) A-C in surface air temperature. Unit: K. (e) Runoff in A run (contours) and C run (tones). Unit: mm d⁻¹.
Figure 5: A-C in surface air temperature in April (contours) and July (tones). Unit: K. Dashed lines denote negative values.

Figure 6: Latitude height section of A-C in temperature at 90°E in April (contours) and July (tones). Unit: W m$^{-2}$. Dashed lines denote negative values.
5. Summary and Future Directions

The effects of the land surface processes in cold regions, such as soil freezing and snow, have attracted an attention in the recent years. Viterbo et al. (1999) show the importance of soil freezing in the numerical weather forecast in winter. Takata and Kimoto (2000) present the impact of soil freezing on the continental-scale seasonal cycle of water and energy particularly in the Asian monsoon region.

The impact of the treatment of snow albedo on the atmosphere is examined using an AGCM. Two AGCM experiments with a prognostic snow albedo (C run) and with a fixed snow albedo at a new snow value (A run) are compared. The change in snow albedo is higher in autumn than in spring, but its impact on the atmosphere is larger in spring than in autumn. That is because the changes in the sub-grid snow fraction is higher in spring than in autumn, thereby affects the energy balance in a grid cell more in spring. Hence the change in upward shortwave radiation is higher in spring, causing the large temperature difference then. The temperature difference is spread over Eurasian and American continents and is extended to the lower troposphere. These results imply that it is important to investigate a proper way to parameterize the sub-grid snow coverage, which is inevitable for large-scale models whose grid size is large.

The land surface models hitherto used in the global climate models have considered only the water and energy exchange between the land and the atmosphere. The recent land surface models incorporate the CO₂ exchange between the plant and the atmosphere. Further, the interaction between the vegetation and the climate will be fully treated in near future. It is pointed out by Numaguti (1999) that the land surface evapotranspiration has a considerable effect on the hydrological cycle in the middle and high latitudes of Eurasian continent. Thus the role of interaction among the atmosphere, the land surface processes including frozen ground, and the vegetation will be the next issue to be investigated in the cold regions hydrological cycle.

6. References


1. GAME Project

The GEWEX Asian Monsoon Experiment (GAME) has been ongoing since 1996, focussing primarily on observational studies at field sites in environments ranging from tropical to the cold Siberia region, to determine land-surface/atmosphere interaction related to Asian monsoon formation (GISP, 1998). A research group was formed for each study region to conduct various kinds of meteorological and hydrological observations related to the water and energy cycle which differs from region to region depending on the regional climatic characteristics. Most of the results have been reported in the Proceedings of the International Conference “GEWEX in Asia and GAME” (GISP, 2001). Another 2-3 years (GAME-II) is planned for synthesis and modelling activities.

2. Objectives of GAME-Siberia

Eastern Siberia was selected as one of the main study regions for GAME for several reasons. Firstly, it is a northern region with extensive snow cover and permafrost, possesses typical surface conditions such as taiga forest and tundra, and is representative of a large area of the Eurasia Continent. Secondly, there is abundant freshwater runoff to the Arctic Ocean which may strongly affect sea ice conditions, ocean circulation and resultant climate conditions in the Arctic and surrounding area. Thirdly, this region is centered in an area of recent intense warming and better understanding is needed on the response of the land surface and possible feedbacks to the climate system.

In order to progress the study, we set the following objectives:

1) Clarify the physical processes of the land-surface/atmosphere interacting system;
2) Clarify the characteristics and variability of the regional energy/water cycle;
3) Determine the climate trend and land-surface change during the past 50 years and evaluate possible feedback processes;
4) Improve and develop models describing the energy/water exchange and atmosphere/land-surface systems;
5) Collect and archive regional ground-based and satellite data; and
6) Establish an observational network for long-term study, and develop hardware.
3. Results of GAME-Siberia

The four main strategies for implementing the GAME-Siberia study were to:

1) select one drainage area for study. The Lena River basin, the eastern most drainage area among the three large Arctic flowing rivers, was the catchment selected;
2) establish three local observation sites for intensive study based on the criteria of land surface condition and climate in the drainage basin. The three sites established were a tundra area facing the Arctic Ocean, flat taiga with low precipitation, and mountain taiga with high precipitation;
3) hold an intensive observation period (implemented in year 2000) for investigating the land-surface/atmosphere interaction and spatial and temporal variability of water/energy fluxes on a regional scale (100 km scale); and
4) involve researchers from various disciplines that could contribute to the understanding of the water/energy cycle in this region.

The study period was 1996-2001, and tight cooperation between Japanese and Russian institutions and scientists lead to a successful study and operation. The main results obtained are as follows:

1) Tundra surface processes: The shallow permafrost table (active layer depth being 50-100 cm) appears to have a rather strong effect on the surface heat budget. Hydrological response (summer evaporation) of the tundra surface is rather stable in comparison to heat supply to the atmosphere (sensible heat). Runoff seems to be regulated by the total winter snow accumulation and heterogeneity of snow distribution due to topography, which is determined by winter climate.

2) Flat taiga surface processes: The seasonal progression of heat/water exchange differs between forest type, such as the most dominant larch and pine. At both sites the influence of snow cover is masked by the canopies. Grass surfaces, in comparison with nearby forested areas, exhibit different seasonal progression of fluxes. In general, the grass cover has lower evaporation and very low sensible heat supply to the atmosphere. This seems to be the result of the shallow permafrost table. Inter-annual variation of evaporation seems to be small in the forest, in contrast to the large inter-annual variability of soil moisture, mainly due to the active function of tree growth. Stable isotope analysis showed that trees pump up water from deeper layers in dry summer years.

3) Land surface - atmosphere interaction at flat taiga: According to aircraft observations of the heterogeneous land surface at a 100 km scale, the distribution of low level (100 m) heat/water fluxes show a complex pattern influenced by characteristics of the boundary layer and local circulation due to the presence of Lena River. A complex land-surface scheme for land-surface/atmosphere interaction needs to be incorporated to evaluate spatial mean fluxes.

4) Precipitation recycling: Stable Isotope analysis shows seasonal change in characteristics of precipitation recycling (weaker in the early summer and stronger in late summer). Seasonal change in the water cycle was detected in these studies.

5) Hydrological model sensitivity tests: Hydrological models applicable to the large drainage area were developed, and rather good results showed that drainage is more sensitive to change in precipitation than to air temperature.
6) Development of an automated year-round observation system: An automated observation system enabled acquisition of year-round meteorological and ground surface data at the three local observation sites.

Other work such as comparison of land hydrology at three local observation sites, application of RAMS Model to the year 2000 Intensive Observation Period (IOP), and satellite analysis of the vegetation changes and land surface conditions is currently being done.

4. Future of GAME

GAME-I (1996-2001) was an observational, data collection phase. GAME-II (2002-2003 (2004)) will be a synthesis phase, directed to crosscut GAME regions with integration studies. The main topics, as discussed by GISP (GAME International Science Panel) in October 2001, will be:

1) Water and Energy Balance (→ WEBS)
2) Land Surface Processes
3) Precipitation Processes
4) Monsoon Processes
5) 4DDA
6) GAIN (Data Archive)
7) Siberia

The Siberia Regional Project is included among the regional projects, since it had a late intensive observation period in year 2000 (other regions had their IOP in 1998).

The observation network of GAME-I will be transferred to CEOP-CAMP, GAME-Tropics II, Frontier and other projects. Main observation networks that need further accumulation of data will be continued and developed.

The next phase for water/energy cycle study in Siberia will be made under the following principles.

1) Synthesis of the regional project and comparison with other regions will be made within the framework of GAME-II.
2) Observational studies and observational systems will be transferred to FORSGC (funded) and other projects (not funded).
3) The Siberian component, in the long-term, will be reorganized under the framework of WCRP-CliC or GEWEX-GHP.

Among the issues for water and energy cycle study in Siberia in the framework of GAME-II is the investigation of the universality of the results obtained from process studies, and transferability of models and concepts (this is the first priority). In order to accomplish this goal, GAME, and at its center GAME-Siberia, hopes to have strong interactions with the Canadian GEWEX component (MAGS) since they are working on similar geographical environments. Both environments have Arctic to sub-Arctic drainage basins, and rivers that flow into the Arctic Ocean; hence, both have common geographical situations and interests. GAME hopes to have
more discussion with the MAGS group regarding collaborative research and further development of the relationship between these projects.

5. References

The purpose of this paper is to provide an overview of the Mackenzie GEWEX Study (MAGS) and to identify some of the key activities that are ongoing and planned for the near future. MAGS has just completed a transition from its first to its second phase. The overall goals of this second phase remain the same as those of the first; namely:

- to understand and model the linked hydrologic-atmospheric system of the Mackenzie River basin,
- to provide tools to predict system response to climate variability and climate change for a variety of needs,
- to enhance the Canadian scientific skill base in hydrology and climatology,
- to contribute to resolution of global issues related to water and climate.

As well as being directed toward national scientific priorities, MAGS is the major Canadian contribution to the Global Water and Energy Cycle Experiment (GEWEX) coordinated by the World Climate Research Program. The objectives of GEWEX are:

- to measure global hydrological cycle and energy fluxes,
- to model global hydrological cycle and its impact on atmosphere, oceans and land surfaces,
- to predict global and regional response of water resources to environmental change,
- to advance observing techniques and data management and assimilation systems.

Although the transition from the first to the second phase of MAGS does not involve any change in the study goals, it does involve a major shift in the research focus (Table 1).

Table 1: The MAGS-1 to MAGS-2 transition.

<table>
<thead>
<tr>
<th>MAGS-1</th>
<th>MAGS-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>1996</td>
<td>1 January 2001</td>
</tr>
<tr>
<td>• Data collection, management and assimilation</td>
<td>• Process integration</td>
</tr>
<tr>
<td>• Atmospheric and hydrologic process studies</td>
<td>• Develop linked hydrologic-atmospheric models</td>
</tr>
<tr>
<td>• Modelling framework and capability</td>
<td>• Apply models and understanding to environmental and social issues</td>
</tr>
</tbody>
</table>

MAGS-1 was a foundation study that was largely concerned with:

- the collection, management and assimilation of data on the Mackenzie River basin,
- conducting basic atmospheric and hydrologic process studies,
- developing a modelling framework and capability.
MAGS-2 builds on this foundation to:

- integrate the results of process studies into a conceptual understanding of water and energy cycling in the Mackenzie River basin,
- develop a physically-based, predictive model of water and energy flow in the hydrologic-atmospheric system of the Mackenzie Basin,
- apply data, understanding and models to environmental and social issues.

The transition to MAGS-2 brings a change in the organization of scientific activities. MAGS-2 has 11 specific objectives that are grouped under 5 theme areas. Each theme has a designated leader who sits as a member of the SC and is responsible for coordinating research activities in each theme area. The themes and objectives are summarized in Table 2.

Table 2: MAGS-2 themes and objectives.

<table>
<thead>
<tr>
<th>THEMES</th>
<th>OBJECTIVES</th>
</tr>
</thead>
<tbody>
<tr>
<td>I. Process studies and integration</td>
<td>1. Extend process studies</td>
</tr>
<tr>
<td></td>
<td>2. Integrate process studies to produce a unified atmospheric-hydrological framework</td>
</tr>
<tr>
<td>II. Scaling of data and processes</td>
<td>3. Parameterization techniques</td>
</tr>
<tr>
<td></td>
<td>4. Bridge temporal and spatial scales</td>
</tr>
<tr>
<td>III. Model development and evaluation</td>
<td>5. Develop a hierarchy of models</td>
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<tr>
<td></td>
<td>6. Improve coupled models</td>
</tr>
<tr>
<td></td>
<td>7. Evaluate model performance</td>
</tr>
<tr>
<td>IV. Prediction and analyses</td>
<td>8. Close the water budgets</td>
</tr>
<tr>
<td></td>
<td>9. Assess responses to climate forcings</td>
</tr>
<tr>
<td>V. Applications and model transfer</td>
<td>10. Application to problems</td>
</tr>
<tr>
<td></td>
<td>11. Transfer of information and models</td>
</tr>
</tbody>
</table>

These themes are associated with a timeline of activities. This is shown in Table 3. Activities have proceeded according to schedule through the first year of MAGS-2. Several new activities are scheduled to be initiated in the year ahead. These are:

- the analysis of process data collected in Phase-2,
- the development of data upscaling procedures,
- the development and validation of new process models.
Other activities are scheduled to be completed by the end of the 2002. These include:
- the integration of process studies into a conceptual foundation for modelling,
- the identification of user needs for model application.

Table 3: The MAGS-2 thematic timeline.

<table>
<thead>
<tr>
<th>Theme</th>
<th>Objective</th>
<th>Components</th>
<th>2001</th>
<th>2002</th>
<th>2003</th>
<th>2004</th>
<th>2005</th>
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<tbody>
<tr>
<td>Processes and Integration</td>
<td>1</td>
<td>Measurements</td>
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<td></td>
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<tr>
<td></td>
<td>2</td>
<td>Integration</td>
<td></td>
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<td></td>
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</tr>
<tr>
<td>Scaling Data and Processes</td>
<td>3</td>
<td>Distributed data</td>
<td></td>
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<td></td>
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<td>Upscaling</td>
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<td>4</td>
<td>Modeling</td>
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<td>Model Development and Evaluation</td>
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<td>CRCM</td>
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<td>CLASS/WATFLOOD</td>
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<td></td>
<td></td>
<td>Process models</td>
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<td></td>
<td>6</td>
<td>Coupled models</td>
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<td></td>
<td>7</td>
<td>Validation</td>
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<td>Prediction and Analysis</td>
<td>8</td>
<td>Meso-scale</td>
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<td>Sensitivity</td>
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<tr>
<td>Applications and Model Transfer</td>
<td>10</td>
<td>User feedback</td>
<td></td>
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<td>Within Canada</td>
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<td>Other CSEs</td>
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</tbody>
</table>

An important new development in the past year was the initiation of thematic workshops to focus on specific scientific activity areas. Two thematic workshops were held in the past year to discuss issues around process integration and scaling. These are summarized below.

Theme I (Process Integration) Workshop
- **Purpose:**
  - Discuss state of modelling capability and process studies at MAGS-1 to 2 transition
  - Explore mechanisms for integrating processes, data and models
- **Major discussion points**
  - Coupled vs linked models
  - To be incorporated into MAGS models
    - Infiltration into frozen ground
    - Permafrost
    - River ice jams
    - Wetlands and lakes
    - Sublimation of blowing snow
- Importance of scale
  - Observations and remote sensing data
  - Parametric representation of processes
- The nature of modelling
  - "Acceptable" levels of error
  - Springboard to future models
- Scientific questions that need to be addressed
  - What physics must be incorporated into models?
  - Are data sets adequate for verification?
  - Utility of case studies (CAGES etc.)?

Theme II (Scaling) Workshop
- Scientific workshop, field trip, and outreach activities
- Scientific objectives:
  - Move forward the MAGS scaling theme
  - Strengths/weaknesses of scaling research
  - Identify specific actions to meet scaling theme objectives
- Actions identified as required by area:
  - Atmospheric studies
    i. Large-scale influences on weather and fast climate in MRB
    ii. Numerical scaling experiments with RCM and MC2 for CAGES cases
    iii. Severe weather climatology incorporating lightning data
  - Surface studies
    i. Automate WATFLOOD parameterization
    ii. Large (i.e. resolvable) lakes interactions
    iii. Test a CLASS parameterization of small (i.e. not resolvable) lakes
    iv. Characterize and assess impact of sub-grid lake size frequency distribution
  - Hydrologic studies
    i. Apply atmospheric scaling experiment output to hydrologic models
    ii. Downscaling coarsely observed surface data including the potential use of statistical-dynamical approaches
    iii. Topological arrangements in modelling linkages of land-surface mosaic
- Other activities and outcomes
  - Seminar on models needed
  - Field trip: “Modellers in the Muskeg”
  - Outreach: community, industry, government
  - Application issues
    i. Small-scale hydropower
    ii. River transport (ice season, water levels)
    iii. Traditional lifestyle sustainability
Some other highlights of the past year included:

- MAGS-1 synthesis article in progress
- MAGS 6th Scientific Workshop, Saskatoon, 10-16 November 2000
- MAGS special issue of *Atmosphere-Ocean* on 1994-95 WY completed
- MAGS special issue of the *Journal of Hydrometeorology* on Canadian GEWEX Enhanced Study (CAGES) under development
- Saskatchewan GEWEX Experiment (SAGE) planned as internationally significant transferability study
- Initial water budget closure at CSE scale

Some planned activities for the year ahead include:

- Workshops and training seminars
  - CAGES and data management
  - Models training workshop
  - WEBS workshop
- MAGS scientific planning meeting (Mar 2002)
- Completion of the CAGES special issue of *Journal of Hydrometeorology*
- Development of a MAGS contribution to WEBS and CEOP international GEWEX projects
- Transferability studies
  - Continuing collaboration with BALTEX and GAPP
  - SAGES model transferability study planning continues

The GEWEX Hydrometeorology Panel (GHP) has proposed a set of criteria for assessing the progress of its Continental Scale Experiments (CSE), of which MAGS is one. The scientific criteria are presented in Table 4. Each criterion is rated as completed (C), progressing (Pr) or beginning (B). Italicized entries have been upgraded in the past year.

<table>
<thead>
<tr>
<th>Criteria</th>
<th>Status</th>
</tr>
</thead>
<tbody>
<tr>
<td>Simulate the diurnal, seasonal, annual and interannual cycles.</td>
<td>Pr</td>
</tr>
<tr>
<td>Close water and energy budgets.</td>
<td>Pr</td>
</tr>
<tr>
<td>Determine and understand climate system variability and critical feedbacks.</td>
<td>Pr</td>
</tr>
<tr>
<td>Demonstrate improvements in predictions of water-related climate parameters.</td>
<td>Pr</td>
</tr>
<tr>
<td>Demonstrate the applicability of techniques and models to other regions.</td>
<td>B-Pr</td>
</tr>
</tbody>
</table>

Table 4: Scientific criteria for CSE assessment.
The technical-logistical criteria are presented in Table 5. Each criterion is rated as functioning (F), initiating (I) or planned (P). Italicized entries have been upgraded in the past year.

Table 5: Technical-logistical criteria for CSE assessment.

<table>
<thead>
<tr>
<th>Criterion</th>
<th>Rating</th>
</tr>
</thead>
<tbody>
<tr>
<td>NWP centre atmospheric and surface data assimilation and estimates of hydro-meteorological properties.</td>
<td>F</td>
</tr>
<tr>
<td>Suitable atmospheric-hydrological models and numerical experimentation and climate change studies.</td>
<td>F</td>
</tr>
<tr>
<td><em>Mechanism for collecting and managing adequate hydrometeorological data sets.</em></td>
<td>F</td>
</tr>
<tr>
<td>Participate in the open international exchange of scientific information and data.</td>
<td>I-F</td>
</tr>
<tr>
<td>Interactions with water resource agencies and related groups to address the assessment of impacts on regional water resources.</td>
<td>I-F</td>
</tr>
<tr>
<td>Evaluation of GEWEX global data products</td>
<td>I</td>
</tr>
<tr>
<td>Contributions to CEOP and transferability databases.</td>
<td>P</td>
</tr>
</tbody>
</table>

It became a tradition through MAGS-1 for an annual statement to be prepared by the chair of the Scientific Committee summarizing activities and progress over the previous year. Following is the summary statement for 2001.

*MAGS-1 has been brought to a successful conclusion and a transition to MAGS-2 has been accomplished. A MAGS-1 synthesis article is being prepared and it will illustrate the realization of our process, model framework and data objectives. MAGS-2 is being launched through specific activities such as focused integration and scaling workshops, synthesis of the CAGES data set and planning of transferability test over the Saskatchewan River basin.*

The slides used for this presentation follow:
Mackenzie GEWEX Study

Dr. Lawrence W. Martz, Ph.D., P. Geo.
Chair, MAGS Science Committee
Professor of Geography, University of Saskatchewan

MAGS is...

- The Mackenzie GEWEX Study.
- A coordinated set of process, remote sensing and modelling studies.
- A collaborative research network of Canadian government and university scientists.
- Intended to improve understanding of the water and energy cycle of the Mackenzie River Basin and of cold regions in general.
Quantify and model the water and energy cycle

GEWEX is...

- The Global Energy and Water Cycle Experiment
- Global Objectives
  - Measure global hydrological cycle and energy fluxes
  - Model the global hydrological cycle and its impact on the atmosphere, oceans and land surfaces.
  - Predict global and regional response of water resources to environmental change.
  - Advance observing techniques and data management and assimilation systems.
The MAGS study area

Approximate outline of Mackenzie Basin
MAGS goals are...

- Understand and model the linked hydrologic-atmospheric system of the Mackenzie Basin
- Provide tools to predict system response to climate variability and climate change for a variety of needs
- Enhance the Canadian scientific skill base in hydrology and climatology
- Contribute to resolution of global issues related to water and climate
MAGS focus is…

“Cold regions” processes that challenge our understanding and ability to model the global water cycle.

Why MAGS?

“It is generally accepted that in the case of water policy and management, our scientific basis for decision-making is woefully inadequate.”

Water Sector: Vulnerability and Adaptation to Climate Change, GCSI and AES (2000), p. 32

Especially true of the Canadian North!
Sensitivity to climate change

Winter temperature trends (1961-1990)

Global impacts
Importance of water cycle

- Environmental changes
  - deltas, wetlands, aquatic ecosystems, permafrost
- National and international water issues
  - water export, diversion, water use agreements
- Resource development
  - hydro power, hydrocarbon, minerals
- Community sustainability
  - water use, traditional lifestyles, agriculture, public health
- Climate feedbacks
  - snow/temperature and gas fluxes
- Freshwater input to the Arctic Ocean

Water cycle elements & issues

- Elements of the water cycle
  - Precipitation
  - Evaporation
  - Snow cover
  - Glacier melt
  - Ice breakup
  - Baseflow
  - Permafrost
- Issues
  - Time lags
  - Feedbacks
  - Local variability
### MAGS-1 & MAGS-2

<table>
<thead>
<tr>
<th>MAGS-1</th>
<th>MAGS-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>1996</td>
<td>1 Jan 2001</td>
</tr>
</tbody>
</table>

- Enhance data collection, management and assimilation
- Improve knowledge of atmospheric and hydrologic cycles
- Develop modelling framework and capability
- Integrate knowledge of atmospheric and hydrological cycles into a unified system
- Develop hierarchy of models for a range of spatial and temporal scales
- Apply improved predictive ability to environmental and social issues

---

### Canadian GEWEX Enhanced Study (CAGES) 1998-99 water year
Atmospheric Studies

GOES-10 IR image
25 Apr 1999 0000 UTC

Surface Hydrology Studies
Coupled hydrologic/atmospheric models

The MAGS Research Network

- Government & university
- Hydrology, climatology, meteorology, remote sensing, geography, engineering...

DIAND
PNR
U of A
U of S
NWRI
McGill U
York U
MSC
U of Toronto
McMaster U
U of Waterloo
U of Waterloo
Application Goals

- Validated, coupled models to provide…
  - understanding and prediction of climate change impacts on natural ecosystems
  - support for social adaptation decisions
  - a scientific basis for planning and policy making
  - improved resource management

- Kit Szeto - Meteorological Service of Canada
  - Atmospheric studies
- Ming-ko Woo - McMaster University
  - Frost Hydrology
List of Participants

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