

# **GLACIOLOGICAL** **DATA**

## **SNOW WATCH '92** **DETECTION STRATEGIES FOR SNOW AND ICE**

### **WORKSHOP ON CRYOSPHERIC DATA** **RESCUE AND ACCESS**

World Data Center A  
for  
Glaciology  
[Snow and Ice]



WDC operated for:  
**U.S. Department of Commerce**  
National Oceanic and Atmospheric Administration  
National Environmental Satellite, Data, and Information Service  
Boulder, Colorado 80303 U.S.A.

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# **GLACIOLOGICAL** **DATA**

**REPORT GD-25**

## **SNOW WATCH '92** **DETECTION STRATEGIES FOR SNOW AND ICE**

*An International Workshop on Snow and Lake Ice Cover  
and the Climate System*

Co-Sponsored by

Canadian Climate Centre, World Meteorological Organization and Institute for Space and Terrestrial Science

Edited by

Roger G. Barry, Barry E. Goodison, and Ellsworth F. LeDrew

A Joint Publication of the  
Earth-Observations Laboratory  
of the Institute for Space and Terrestrial Science and the  
World Data Center A for Glaciology (Snow and Ice)

## **WORKSHOP ON CRYOSPHERIC DATA** **RESCUE AND ACCESS**

Edited by

Robert G. Crane

WDC operated for:

**U.S. Department of Commerce**

National Oceanic and Atmospheric Administration  
National Environmental Satellite, Data, and Information Service  
Boulder, Colorado 80303 U.S.A.

**August 1993**

## DESCRIPTION OF THE WORLD DATA CENTER SYSTEM<sup>1</sup>

The World Data Centers (WDCs) were established in 1957 to provide archives for the observational data resulting from the International Geophysical Year (IGY). In 1958 the WDCs were invoked to deal with the data resulting from the International Geophysical Cooperation 1959, the one-year extension of the IGY. In 1960, the International Council of Scientific Unions (ICSU) Comite International de Geophysique (CIG) invited the scientific community to continue to send to the WDCs similar kinds of data from observations in 1960 and following years, and undertook to provide a revised *Guide to International Data Exchange* for that purpose. In parallel the CIG inquired of the IGY WDCs whether they were willing to treat the post-IGY data; with few exceptions, the WDCs agreed to do so. Thus the WDCs have been serving the scientific community continuously since the IGY, and many of them archive data for earlier periods.

In November 1987 the International Council of Scientific Unions (ICSU) Panel on World Data Centers prepared a new version of the *Guide to International Data Exchange*, originally published in 1957, and revised in 1963, 1973 and 1979. The new publication, *Guide to the World Data Center System, Part 1, The World Data Centers (General Principles, Locations and Services)*, was issued by the Secretariat of the ICSU Panel on World Data Centers. This new version of the *Guide* contains descriptions of each of the twenty-seven currently operating disciplinary centers, with address, telephone, telex, and contact persons listed. The reader is referred to the new *Guide* for descriptions of the responsibilities of the WDCs, the exchange of data between them, contribution of data to WDCs, and the dissemination of data by them. The WDCs for Glaciology are listed below.

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The following organization provides international data services including data analyses and preparation of specialized data products. It merges the previous activity of the Permanent Service on the Fluctuations of Glaciers and the Temporary Technical Secretariat for World Glacier Inventory. These activities are not part of the WDC system but the center cooperates with WDCs in the discipline. Users wishing assistance in seeking data or services from this group may contact an appropriate WDC.

### **World Glacier Monitoring Service (WGMS)**

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<sup>1</sup>Adapted from *Guide to the World Data Center System. Part 1. The World Data Centers (General Principles, Locations and Services)*. International Council of Scientific Unions. Panel on World Data Centers, November 1987, 91pp.

## Foreword

The *Snow Watch '92* workshop, sponsored by the Canadian Climate Centre, the World Meteorological Organization and the Institute for Space and Terrestrial Science, University of Waterloo, was held at Niagara-on-the-Lake, Ontario, 29 March – 1 April 1992. This third meeting of the informal Snow Watch group met to review strategies for detection of changes in global snow and lake ice cover as climate system indicators. The reports and recommendations of the previous meetings, on which the 1992 workshop build, are published in *Glaciological Data, Report* GD-11 (Snow Watch '80) and GD-18 (Snow Watch '85). The working groups examined previous recommendations and noted areas where progress had taken place. Areas where progress was absent received special attention and, wherever possible, the 1992 recommendations have been addressed to specific organizations for their consideration.

WDC-A for Glaciology wishes to thank Dr. E.F. LeDrew, University of Waterloo, and his staff for arranging the meeting. Special thanks are due to Kathleen Lamothe for coordinating and preparing the report for publication.

This issue also contains the report of a Workshop on Cryospheric Data Rescue and Access held at Pennsylvania State University, 11-12 May 1993, and convened by Dr. Roger G. Crane of the Earth System Science Center. The workshop was supported by NOAA's program of Earth System Data and Information Management (ESDIM). The twenty-one workshop participants developed guidelines for selecting data required in support of observational studies of modeling the cryosphere, focussing on snow cover, sea ice, ice sheets and glaciers. Candidate high-priority data sets for these areas are identified in the report, together with data considered at risk and in need of rescue. This meeting was the first workshop in support of the NOAA ESDIM implementation plan and the Data Center thanks Dr. Crane for arranging the meeting and coordinating the report so promptly, and Dave Clark, NOAA/National Geophysical Data Center, for his assistance in arranging support for the Workshop.

Roger G. Barry  
Director  
National Snow and Ice Data Center  
WDC-A for Glaciology

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# SNOW WATCH '92

## DETECTION STRATEGIES FOR SNOW AND ICE

### AN INTERNATIONAL WORKSHOP ON SNOW AND LAKE ICE COVER AND THE CLIMATE SYSTEM

The Pillar and Post Inn  
Niagara-on-the-Lake, Ontario

Co-Sponsored by  
Canadian Climate Centre  
World Meteorological Organization  
Institute for Space and Terrestrial Science

## AGENDA

### SUNDAY, MARCH 29

3:00 PM to Dinner: Registration  
6:30 PM Dinner

### MONDAY, MARCH 30

7:00 AM Breakfast

9:00 AM

INTRODUCTION AND WELCOME

CHAIR: DR. B. GOODISON

Canadian Climate Centre Welcome:

World Meteorological Organization Welcome:

National Snow and Ice Data Centre Welcome:

Previous Snow Watch Meetings:

Logistics:

Introduction of Attendees

Dr. D. K. Dawson

Mr. G. McKay

Dr. R. G. Barry

Dr. G. Kukla

Dr. E. F. LeDrew

Dr. D. Robinson:

RECENT VARIATIONS IN NORTHERN HEMISPHERE SNOW COVER

Dr. C. Ropelewski:

NORTHERN HEMISPHERE SNOW COVER AND TEMPERATURE PATTERNS IN THE 1980'S

### Break

CHAIR: DR. R.G. BARRY

Dr. J. Walsh:

TEMPERATURE VARIATIONS IN NORTHERN HIGH LATITUDES, 1960-1990

Dr. P. Groisman:

INFERENCES OF THE NORTH AMERICAN SNOWFALL AND SNOW COVER  
WITH RECENT GLOBAL TEMPERATURE CHANGES

Dr. R. Armstrong:

DETECTION OF FLUCTUATIONS IN GLOBAL SNOW COVER  
USING PASSIVE MICROWAVE REMOTE SENSING



12:00 PM Lunch

CHAIR: MS. A. WALKER

Dr. E. Kuusisto:

LAKE ICE OBSERVATIONS IN THE 19TH AND 20TH CENTURY - ANY MESSAGE FOR THE 21ST?

Dr. R.G. Barry:

LAKE ICE MONITORING FOR CLIMATE CHANGE DETECTION

Dr. D. Hall:

ANALYSIS OF LAKE ICE AS A CLIMATE INDICATOR USING REMOTE SENSING

Break

CHAIR: DR. E.F. LEDREW

Dr. D. Versegby:

SNOW MODELLING IN GENERAL CIRCULATION MODELS

Dr. G. Kukla:

SNOW AND ICE DURING CLIMATE CHANGE: WHAT TO WATCH FOR

Dr. G. MacNeil:

AN APPROACH TO THE CHANGE-POINT PROBLEM

5:30 Reception and tour - Hillebrand Estates Winery

7:30 pm Dinner

TUESDAY, MARCH 31

7:00 AM Breakfast

CHAIR: DR. D. ROBINSON

15 minute overviews:

Dr. K. Dewey:

SOUTH AMERICAN SNOW COVER DATA

Dr. D. Yang:

RESEARCH AT LANZHOU INSTITUTE (CHINA) ON SNOW COVER/CLIMATE

Dr. C. Benson:

RESEARCH ON SEASONAL AND PERENNIAL SNOW OF ALASKA

Mr. D. Barber and Dr. E. LeDrew:

MODELLING MICROWAVE INTERACTION WITH A SEASONALLY VARIABLE  
SNOW COVER: THE SIMMS EXPERIMENT

Break

CHAIR: DR. J. WALSH

Ms. M. Hughes:

CREATING TEMPORALLY COMPLETE SNOW COVER RECORDS USING OBSERVED  
AND MODELED INPUT

Dr. D. Legates:

REMOVING BIASES FROM RAIN AND SNOWGAGE MEASUREMENTS

Dr. M. Serreze:

DETECTION OF SPRING SNOW MELT OVER ARCTIC SEA ICE

Dr. R. Leconte:

POTENTIAL OF SAR FOR MONITORING SNOWPACKS

Ms. A. Walker:

PASSIVE MICROWAVE FOR SNOW WATER EQUIVALENT AND LAKE ICE

Mr. W. Skinner:

ANALYSIS OF CONVENTIONAL LAKE ICE

12:00 PM Lunch

CHAIR: MR. G. MCKAY

Dr. R.G. Barry and Dr. R. Armstrong (NSIDC):

REVIEW OF RECOMMENDATIONS OF SNOW WATCH '85

Dr. B. Goodison:

DATA NEEDS FOR DETECTION OF CLIMATE CHANGE: AN OPEN CHALLENGE

WORKING GROUPS:

- 1/ Remote Sensing
- 2/ Statistical Procedures
- 3/ Lake Ice Indicators

CHAIR: DR. M. SERREZE

PLENARY SESSION

6:30 pm Dinner

WEDNESDAY, APRIL 1

7:00 AM Breakfast

CONTINUATION OF WORKING GROUPS

- 4/ Data Management
- 5/ Modelling
- 6/ Snow Cover Indicators

CHAIR: DR. K. DEWEY

PLENARY SESSION

12:30 PM Luncheon

Guest: The Honourable Shirley Martin,  
Minister of State for Transport,  
Government of Canada

WRAP-UP DISCUSSION

3:30 PM DEPARTURE

## ISSUES TO BE CONSIDERED DURING THE WORKSHOP

### REMOTE SENSING

What are the major measurement, monitoring and detection tools for snow and lake ice?  
What are the limitations for implementing their use?  
Are there data management problems?  
Are all the data sources easily available?  
What has to be done to ensure effective development of algorithms and their application?  
What are the priorities in remote sensing/cryosphere for the next 5 years (pre-EOS)?

### STATISTICAL PROCEDURES

What are the statistical tools which may be applicable to snow and ice analyses?  
What statistical tools are available for analyzing time series derived from remote sensing products?  
What are the impediments to using appropriate statistical procedures in the assessment of change and detection?  
Can statistical significance ever be ascribed to cryospheric changes?  
What are the priorities in applying statistical procedures to cryosphere problems during the next 5 years?

### LAKE ICE INDICATORS

What data sets exist to study lake ice change?  
What is the role of remote sensing?  
How frequent do data have to be acquired, processed and archived to define freeze-up and break-up?  
What is the current status of national, regional and global data management?  
What are the priorities in lake ice monitoring for the next 5 years?  
How do we define freeze-up and break-up for remote sensing information?

### DATA MANAGEMENT

What is the current status of national, regional and global data management for lake ice and snow cover?  
What are the priorities in data management for the next 5 years?  
How should these priorities be achieved?  
What joint agency co-operation and effort are required to achieve a data base for climate change and detection?

## MODELLING

What are the scale issues (temporal and spatial) in the modelling of snow/lake ice/climate interactions?

What are the issues in the GCM modelling of snow processes?

What is required for the validation of GCM outputs of snow cover?

Can remote sensing make a contribution in the modelling of snow and ice processes?

How important are snow cover and lake ice in climate models?

Are they adequately parameterized?

What are the priorities for research and action in the next 5 years?

## SNOW COVER INDICATORS

What are the important elements of snow cover that should be analyzed for climate change and detection?

What conventional data should be extracted and archived and analyzed?

What is the role of remote sensing?

Are the data acquisition, processing and management schemes adequate?

What are the priorities for research and action for the next five years?

# Monitoring Northern Hemisphere Snow Cover

---

*David A. Robinson  
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## Abstract

Accurate information on snow cover is essential for understanding details of climate dynamics and climate change. It is critical that snow observations be as lengthy and geographically extensive as possible. Data on snow extent and the physical characteristics of continental snowpacks are gathered from ground sites and aircraft or satellite platforms. Here, these means of observing snow cover are reviewed, recent endeavors to improve or enhance data collection are discussed, and some results of short and long-term monitoring efforts are presented. Included are monthly continental and hemispheric snow areas derived from shortwave satellite data using a new consistent methodology, and a discussion of recent deficits in spring snow cover. Also, a comparison of these hemispheric snow areas with estimates of less extensive cover derived from satellite microwave data is presented, and progress in developing regional algorithms to extract more accurate snow information from satellite microwave returns is discussed. Continental snow cover is emphasized, however a discussion of snow on top of arctic sea ice is included. The latter focuses on a new ten-year time series of spring and summer snow conditions in the Arctic Basin. Finally, recommendations are made for ongoing monitoring efforts for the remainder of this decade and beyond, and for retrospective studies covering all or a portion of the past century.

## Introduction

Across the middle and high latitudes of the Northern Hemisphere, the impact of snow on humans and the environment is considerable. Falling snow or snow lying on the ground or on ice influence hydrologic, biologic, chemical and geologic processes at and near the surface of the Earth. Snow exerts an impact on activities as diverse as engineering, agriculture, travel, recreation, commerce, and safety. Empirical and modelling studies also show snow cover to have an influential role within the global heat budget, chiefly through its effect of increasing surface albedo (Berry, 1981; Walsh et al., 1985; Robinson & Kukla, 1985; Kukla et al., 1986; Barnett et al., 1989). Global models of anthropogenically-induced climate change suggest enhanced warming in regions where snow cover is currently ephemeral (Manabe & Wetherald, 1980; Hansen et al., 1984; Dickinson et al., 1987). For this reason, snow cover has been suggested as a useful index for detecting and monitoring such change (Barry, 1985; Schlesinger, 1986).

Accurate information on snow cover is essential for understanding details of climate dynamics and climate change. It is also critical that snow observations be as lengthy and geographically extensive as possible. Snow data are gathered from ground sites and aircraft or satellite platforms. There are advantages and liabilities to each of these sources that are related to their accuracy and coverage. Here, these means of observing snow cover are reviewed, recent endeavors to improve or enhance data collection are discussed, and some results of short and long-term monitoring efforts are presented. Emphasis is on continental snow cover, however a discussion of snow on top of arctic sea ice is also included. Finally, recommendations are made for ongoing monitoring efforts for the remainder of this decade and beyond, and for retrospective studies covering the past century.

## Continental Snow Cover

### *Surface-Based Observations*

Surface-based snow cover data are mainly gathered from observing stations on a once per day basis. The general practice is to record the average depth of snow lying on level open ground having a natural surface cover. At primary stations, the water equivalent of the snowpack may also be measured. In some regions, snow courses have been established, where snow depth, water equivalent and perhaps other pack properties are measured along prescribed transects across the landscape. Observations are often limited to once per month and the number of courses is extremely limited in North America. More frequent and abundant course data are gathered in the Commonwealth of Independent States, although to date these data are not digitized and are generally unavailable in any form (A. Krenke, per. commun.). The remainder of this discussion deals with station (point) observations.

Current station observations of snow cover are of a sufficient density for climatological study in the lower elevations of the middle latitudes. Elsewhere, data are spotty at best. There is no hemispheric snow cover product based entirely on station reports. The U.S. Air Force global snow depth product relies heavily on surface-based observations as input into a numerical model that creates daily charts with global coverage (Hall, 1986). Disadvantages include having to rely on extrapolations and climatology in data-sparse regions (McGuffie & Robinson, 1988). Recently, an improved interpolation scheme incorporating horizontal and vertical components has been added to the model (Armstrong & Hardman, 1991). A method to determine depth using microwave data in data sparse regions has also been added to reduce the reliance on climatology.

There have been a number of regional snow cover products over the years that are based on station data. Of greatest longevity are the WWCB charts that have been produced since 1935. These, and daily U.S. National Oceanic and Atmospheric Administration (NOAA) charts, are produced for the conterminous United States solely from first order station observations, thus neither has a particularly high resolution. Since December 1983, the position of the snow line has not been plotted on the WWCB charts. Only point data for those stations reporting snow cover has been presented. This is a major limitation, as the charts give no indication as to where the snow-free stations are located.

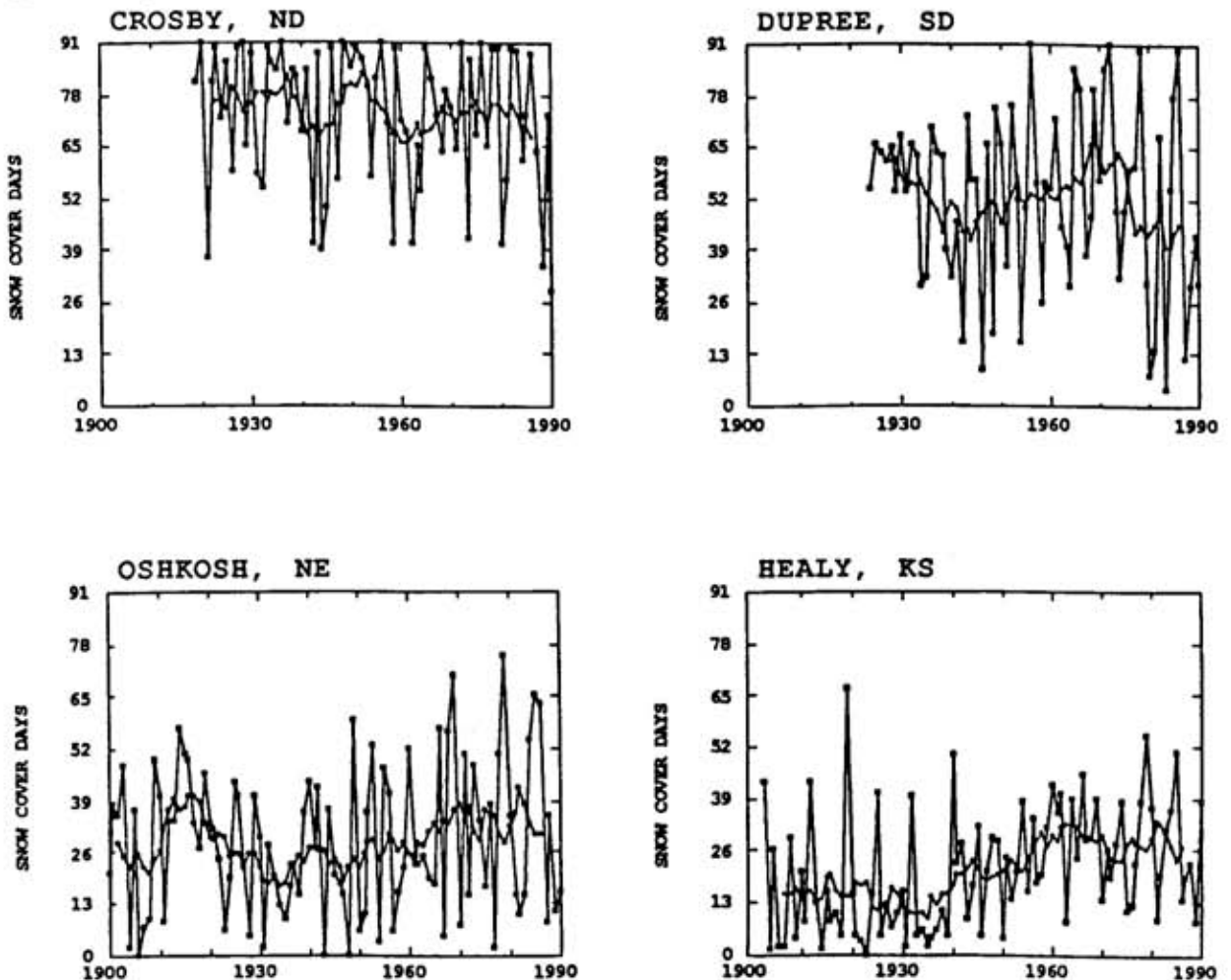
Observations from the primary stations used in the preparation of WWCB and other regional charts are among the most complete and accurate over the long term. However, the consistency of first order station observations suffers from the numerous station relocations from cities to airports in the middle part of this century (particularly in the U.S.). Potential influences of urbanization on the depth and duration of snow on the ground also must be considered. To date, no comprehensive study has been done on either of these subjects.

In a number of countries, there are numerous stations with relatively complete records of snow extending back fifty years or more. Until recently, most data have remained unverified and disorganized (Robinson, 1989). As a result, few studies have dealt with long-term trends or low-frequency fluctuations of snow over even small regions (Arakawa, 1957; Manley, 1969; Jackson, 1978; Pfister, 1985). Through the cooperative efforts of a number of scientists and data centers, this situation has begun to be rectified. Examples include the exchange of data through the US/USSR Bi-Lateral Environmental Data Exchange Agreement and between the Lanzhou Institute of Glaciology and Geocryology and Rutgers University. These and other data are in the process of being quality

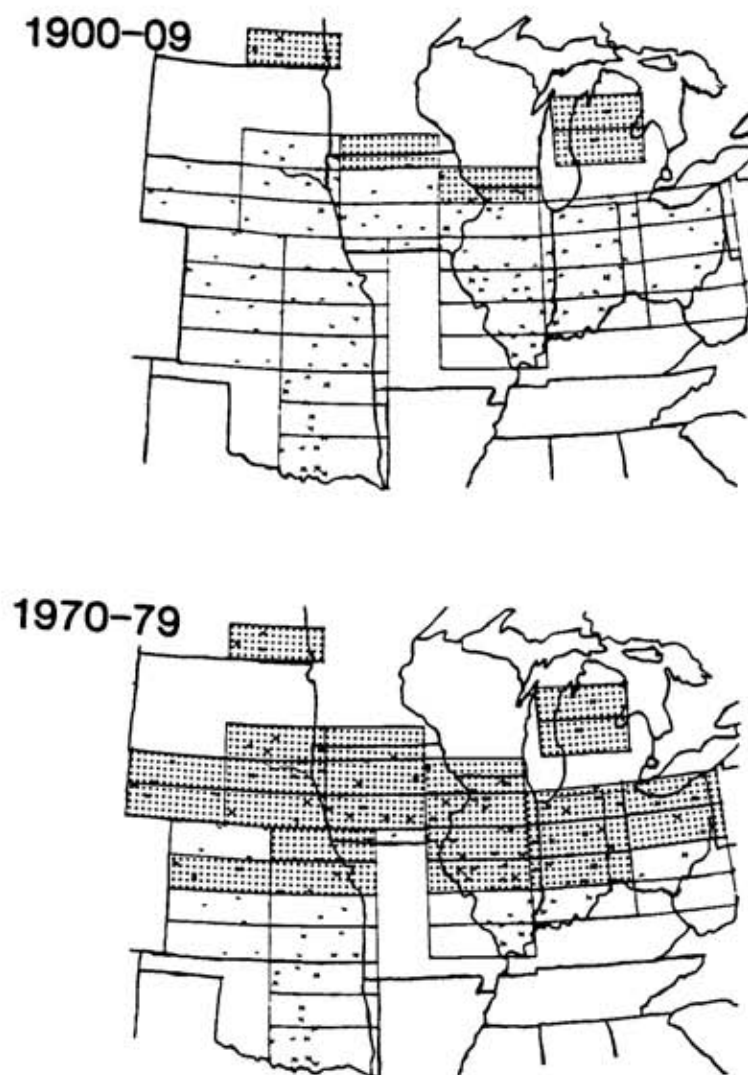


controlled, and routines to fill in gaps in snow cover records are being developed (Robinson, 1988; Hughes & Robinson, 1992).

These data have begun to be analyzed for several regions. For example, marked year-to-year variability in snow cover duration is recognized over the course of this century in the U.S. Great Plains (fig. 1). A broader analysis of this region and areas further east, using data from 146 stations, shows notable decadal variations in snow extent. These stations, all with records back to at least 1910, are grouped into  $1^\circ$  latitude  $\times$   $4^\circ$  longitude divisions to account for the inhomogeneous spacing of the stations and occasional data gaps in the network. Division values are an average of all available reports, which range from one to eight depending on the day and the region. Of the past nine decades, the January snowline was at its southernmost position in the 1970s and was at its northernmost location, roughly three or four degrees of latitude poleward of the 1970s line, in the 1900s (fig. 2). A division is considered to be snow covered in a given decade when more than half of the days have a cover of  $\geq 2.5$  cm for at least five Januaries. A decade-by-decade count of the number of divisions meeting this criterion shows the first five decades of this century to have less January snow cover than the most recent four (fig. 3). This is particularly notable in the central Great Plains.

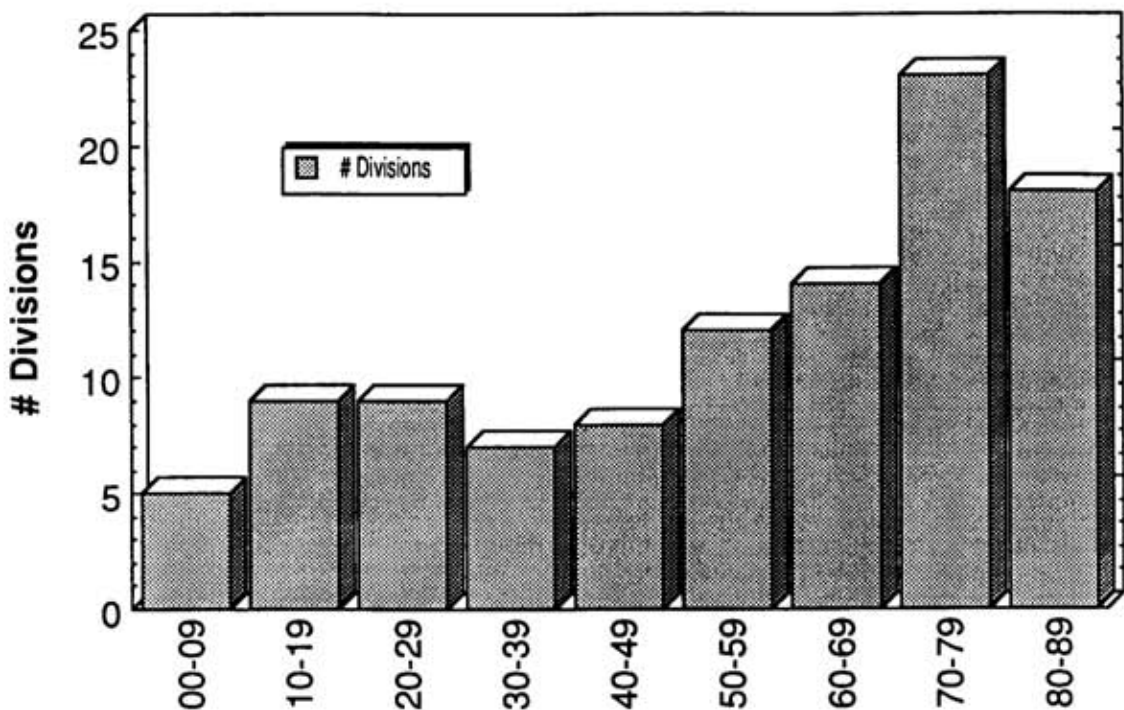


**Figure 1.** Time series of winter days with  $\geq 2.5$  cm of snow cover at several stations on the U.S. Great Plains. Also shown smoothed with a nine-point binomial filter, with only those points plotted where nine years are available for filtering (e.g., plotted year  $\pm 4$  yr).



**Figure 2.** January snow cover over the central U.S. from 1900-09 and 1970-79. Study divisions with five or more years with more than half of the days in the month with a snow cover of  $\geq 2.5$  cm are stippled. Study stations are marked with Xs.





**Figure 3.** Decadal summary of study divisions in the central U.S. (cf. fig. 2) with more than half of the days in January having a snow cover of  $\geq 2.5$  cm in at least five years. Pre-1980 data from the station network, 1980-89 data from NOAA weekly snow charts.

### *Aircraft-Based Observations*

Observations of snow cover from aircraft are limited in both space and time, and when made tend to be for specific investigations. One ongoing program that provides useful information for water supply and flood forecasts over portions of the U.S. and Canada is conducted by the Office of Hydrology of the U.S. National Weather Service. Sensors on board low flying aircraft measure natural terrestrial gamma radiation, from which the extent and mean areal water equivalent of a snowpack is inferred. Knowing the background (snowfree) radiation along a flight line permits subsequent estimates of snow water equivalent by calculating the attenuation of the radiation signal. Flight lines are typically 16 km long and 300 m wide, covering an area of approximately 5 km<sup>2</sup>. A network of 1400 lines covering portions of 25 states and 7 Canadian provinces is currently being flown under this program. The principal sources of error using this methodology are incomplete or inaccurate information on mean soil moisture and characteristics of the air mass along the flight line, and radiation counting limitations (Carroll & Schaake, 1983). Underestimates occur if substantial amounts of forest biomass exist in the region or if the snow cover along the line is uneven (Carroll & Carroll, 1989 a & b).

### *Space-Based Observations*

Regional and hemispheric snow extent is monitored using data recorded in shortwave (visible and near-infrared) and microwave wavelengths by sensors on board geostationary and polar orbiting satellites. Several snow cover products are produced using these data.

## *Shortwave Charting*

Shortwave data provide continental coverage of snow extent at a relatively high spatial resolution. Snow is identified by recognizing characteristic textured surface features and brightnesses. Information on surface albedo and percent snow coverage (patchiness) is also gleaned from the data. Shortcomings include, 1) the inability to detect snow cover when solar illumination is low or when skies are cloudy, 2) the underestimation of cover where dense forests mask the underlying snow, 3) difficulties in discriminating snow from clouds in mountainous regions and in uniform lightly-vegetated areas that have a high surface brightness when snow covered, and 4) the lack of all but the most general information on snow depth (Kukla & Robinson, 1981; Dewey & Heim, 1982).

## *NOAA Snow Charts*

In 1966, NOAA began to map the snow and ice areas in the Northern Hemisphere on a weekly basis (Matson et al., 1986). That effort continues today, and remains the only such hemispheric product. NOAA charts are based on a visual interpretation of photographic copies of shortwave imagery by trained meteorologists. Up to 1972, the subpoint resolution of the meteorological satellites commonly used was around 4 km. Beginning in October 1972, the Very High Resolution Radiometer (VHRR) provided imagery with a spatial resolution of 1.0 km, which in November 1978, with the launching of the Advanced VHRR (AVHRR), was reduced slightly to 1.1 km. Charts show boundaries on the last day that the surface in a given region is seen (fig. 4). Since May 1982, dates when a region was last observed have been placed on the charts. An examination of these dates shows the charts to be most representative of the fifth day of the week.

It is recognized that in early years the snow extent was underestimated on the NOAA charts, especially during Fall. Charting improved considerably in 1972 with the deployment of the VHRR sensor, and since then charting accuracy is such that this product is considered suitable for continental-scale climate studies (Kukla & Robinson, 1981).

Despite the shortwave limitations mentioned earlier, the NOAA charts are quite reliable at many times and in many regions. These include regions where, 1) skies are frequently clear, commonly in Spring near the snow line, 2) solar zenith angles are relatively low and illumination is high, 3) the snow cover is reasonably stable or changes slowly, and 4) pronounced local and regional signatures are present owing to the distribution of vegetation, lakes and rivers. Under these conditions, the satellite-derived product will be superior to charts of snow extent gleaned from station data, particularly in mountainous and sparsely inhabited regions. Another advantage of the NOAA snow charts is their portrayal of regionally-representative snow extent, whereas charts based on ground station reports may be biased due to the preferred position of weather stations in valleys and in places affected by urban heat islands, such as airports.

The NOAA charts are digitized on a weekly basis using the National Meteorological Center Limited-Area Fine Mesh grid. This is an 89 x 89 cell northern hemisphere grid, with cell resolution ranging from 16,000 km<sup>2</sup> to 42,000 km<sup>2</sup>. If a cell is interpreted to be at least fifty percent snow covered it is considered to be completely covered, otherwise it is considered to be snow free.

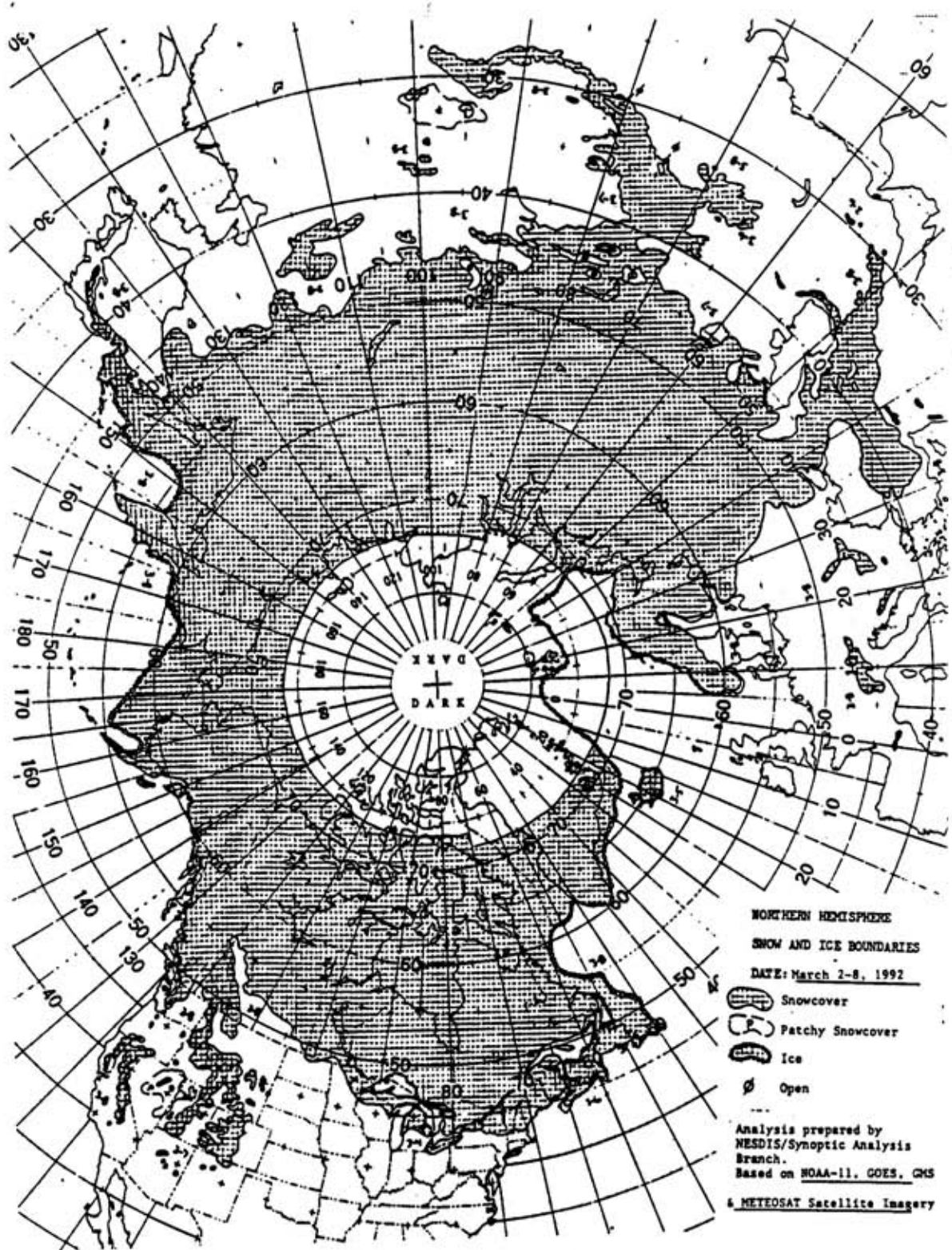
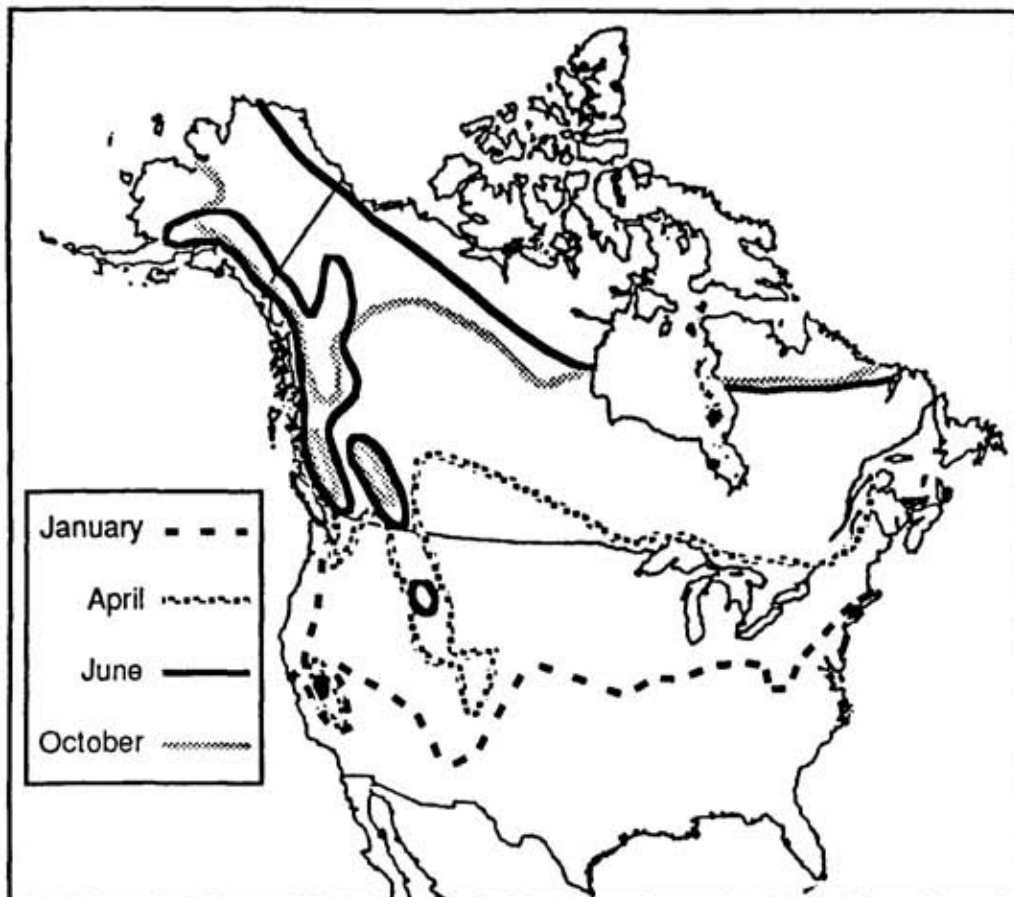


Figure 4. NOAA snow chart for March 2-8, 1992.

Figure 5 shows the mean position of the North American snow line for four months of the year. Snowlines are derived from NOAA digital data by calculating the percentage of the time each digitized cell is recognized as snow covered. To acquire accurate information on the area of land covered with snow for a particular time in a specific year requires a more sophisticated approach. A new routine to calculate monthly snow areas from NOAA data is presented here. The development of this routine follows the discovery of a major inconsistency in the manner in which NOAA has calculated monthly snow cover areas (Robinson et al., 1991). Prior to 1981, NOAA calculated continental areas from monthly summary charts, which consider a cell to be snow covered if snow is present on two or more weeks during a given month (Dewey & Heim, 1982). Since 1981, NOAA has produced monthly areas by averaging areas calculated from weekly charts. A comparison of these two methodologies shows areas computed using the monthly approach to be from several hundred thousand to over three million square kilometers greater than those calculated using weekly areas. The offsets are not consistent. Also contributing to the problem are 53 cells (covering  $1.8 \times 10^6 \text{ km}^2$ ) not considered consistently in the area calculations throughout the period of record. In 1981, NOAA changed their land mask, in the process eliminating 26 cells from consideration of being snow covered (categorizing them as water), while 27 others began to be examined. As discussed below, neither of the NOAA masks is accurate; both fail to accurately identify all cells, and only those cells, at least half covered by land.



**Figure 5.** Mean North American snowlines for October, January, April and June. Lines are derived from NOAA weekly snow charts covering the interval from January 1971 through June 1990. The mean snowline is defined as the isoline denoting a fifty percent frequency of snow cover.



Our new, consistent methodology (Rutgers Routine) calculates weekly areas from the digitized snow files and weights them according to the number of days of a given week that fall in the given month. A chart week is considered to center on the fifth day of the published chart week (cf. above). No weighting has been employed in either of the NOAA routines.

In addition, a definitive land mask has been developed using digital map files analyzed on a geographic information system (GIS). The percentage of land in each of the 7921 NMC grid cells is calculated using the National Geophysical Data Center's five minute resolution ETOPO5 file as the primary data source. As this file does not include large interior lakes, the Navy Fleet Numerical Oceanography Center's 10 minute resolution Primary Terrain Cover Types file is used to properly account for these water bodies. Some 48 cells polewards of approximately 30°N which had been considered land in the pre 1981 NOAA and/or the 1981 to present NOAA mask are actually predominantly water covered (< 50% land). Conversely, 54 land cells are found to have been considered water on one or both NOAA masks. Those cells falling under the latter require a first-time analysis to determine whether they might be snow covered. This is accomplished by selecting nearest representative land cells (cells which NOAA has continuously charted as land) and assigning their snow status to the "new" land cells. Spot checks of a number of hard copy weekly charts prove this to be an adequate approach.

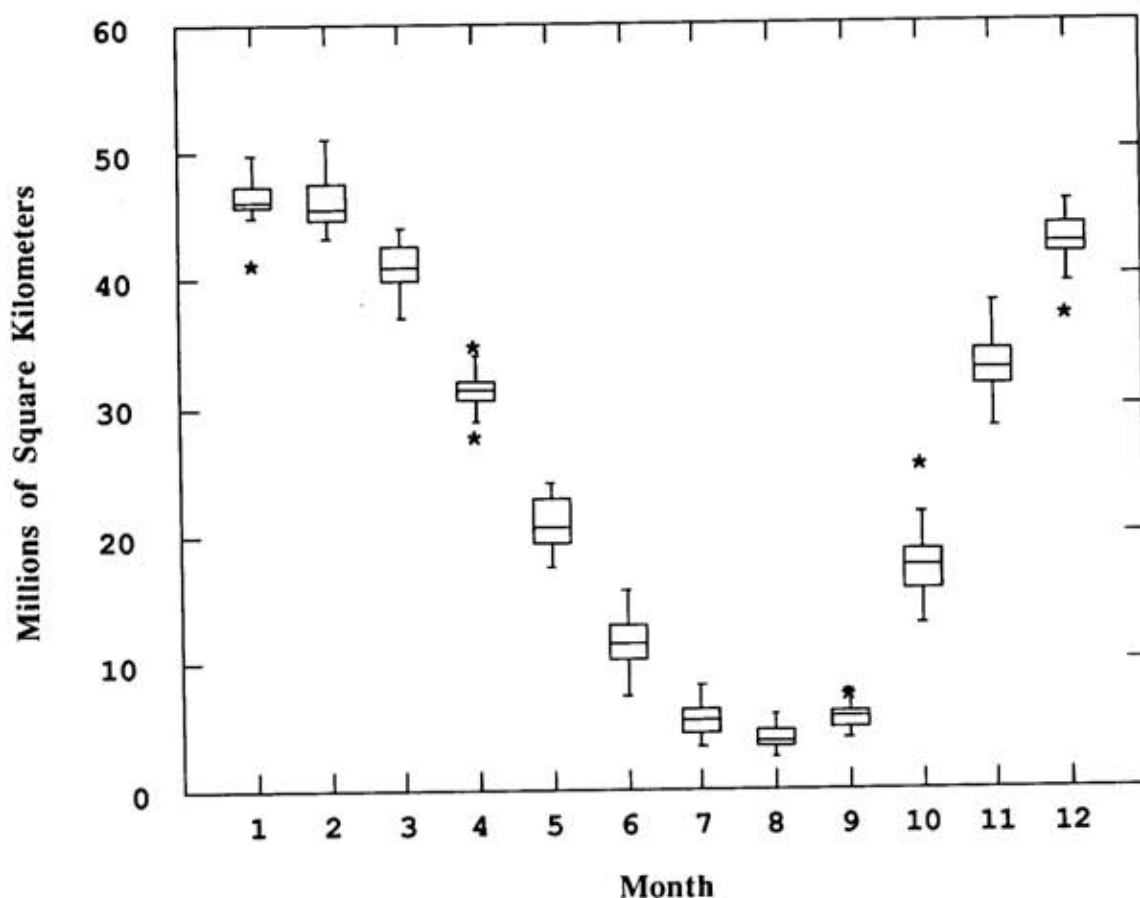
#### ***Continental Snow Cover From NOAA Charts: 1972-1991***

According to values generated using the Rutgers Routine, the extent of snow cover over northern hemisphere lands is greatest in January. On average, some 46.5 million km<sup>2</sup> of Eurasia and North America are snow covered in this month, with February a close second with an average of 46.0 million km<sup>2</sup> (table 1, fig. 6). August has the least cover, averaging 3.9 million km<sup>2</sup>, most of this being snow on top of the Greenland ice sheet. The past two decades of monthly data are close to normally distributed, and monthly standard deviations range from 0.9 million km<sup>2</sup> in August to 2.9 million km<sup>2</sup> in October. The annual mean cover is 25.5 million km<sup>2</sup> with a standard deviation of 1.1 million km<sup>2</sup>. The snowiest year was 1978 with a mean of 27.4 million km<sup>2</sup>, with 1990 the least snowy at 23.2 million km<sup>2</sup>.

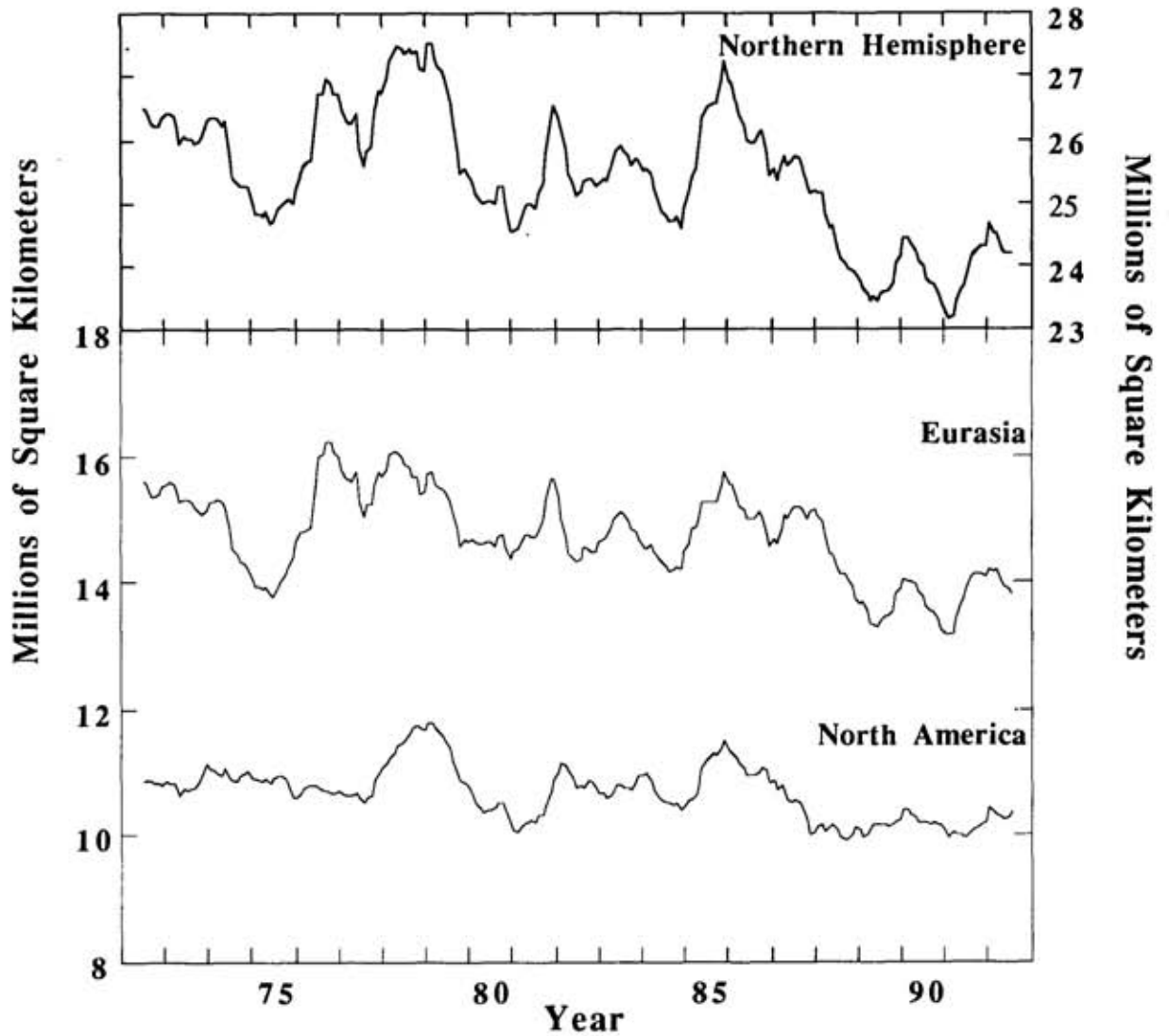
**Table 1.** Monthly and annual snow cover (million km<sup>2</sup>) over northern hemisphere lands during the period January 1972 through May 1992. Areas are calculated using the Rutgers Routine.

	<b>Maximum (yr)</b>	<b>Minimum (yr)</b>	<b>Mean</b>	<b>Median</b>	<b>Std. Dev.</b>
Jan	49.8 (1985)	41.7 (1981)	46.5	46.1	1.8
Feb	51.0 (1978)	43.2 (1990,92)	46.0	45.6	2.0
Mar	44.1 (1985)	37.0 (1990)	41.0	40.8	1.9
Apr	35.3 (1979)	28.2 (1990)	31.3	31.4	1.8
May	24.1 (1974)	17.4 (1990)	20.8	20.6	1.9
Jun	15.6 (1978)	7.3 (1990)	11.6	11.4	2.1
Jul	8.0 (1978)	3.4 (1990)	5.3	5.5	1.2
Aug	5.7 (1978)	2.6 (1988,89,90)	3.9	3.8	0.9
Sep	7.9 (1972)	3.9 (1990)	5.6	5.6	1.1
Oct	26.1 (1976)	13.0 (1988)	17.6	17.5	2.9
Nov	37.9 (1985)	28.3 (1979)	33.0	32.8	2.3
Dec	46.0 (1985)	37.5 (1980)	42.5	42.7	2.3
Annual	27.4 (1978)	23.2 (1990)	25.5	25.4	1.1

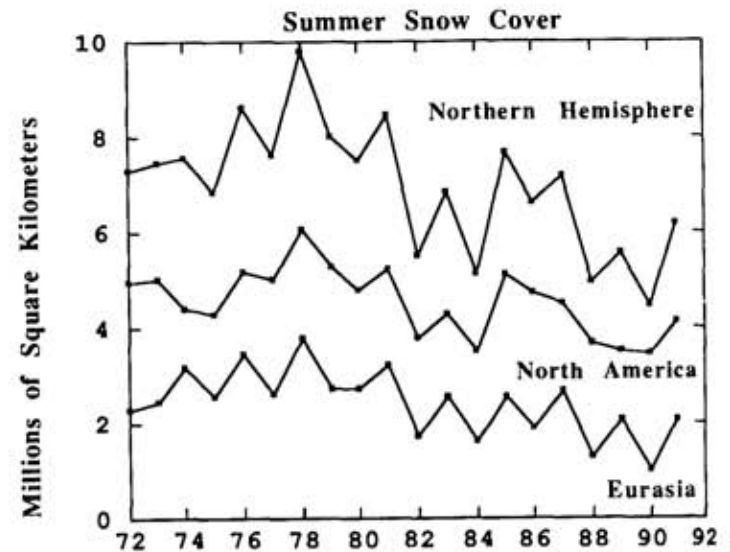
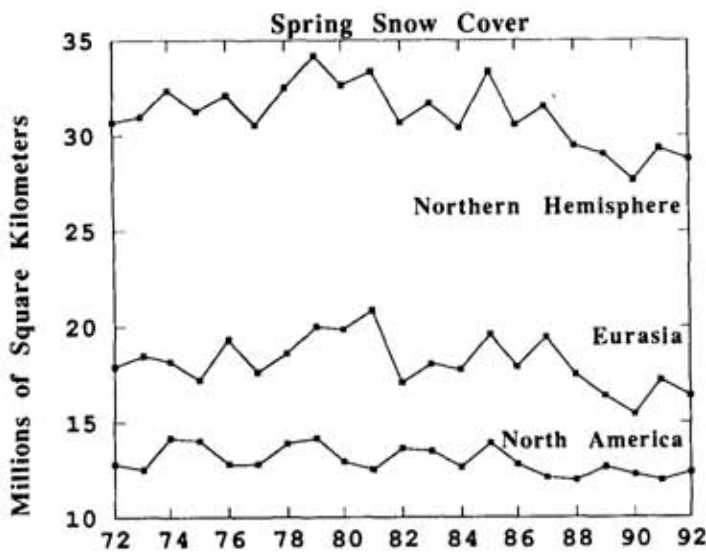
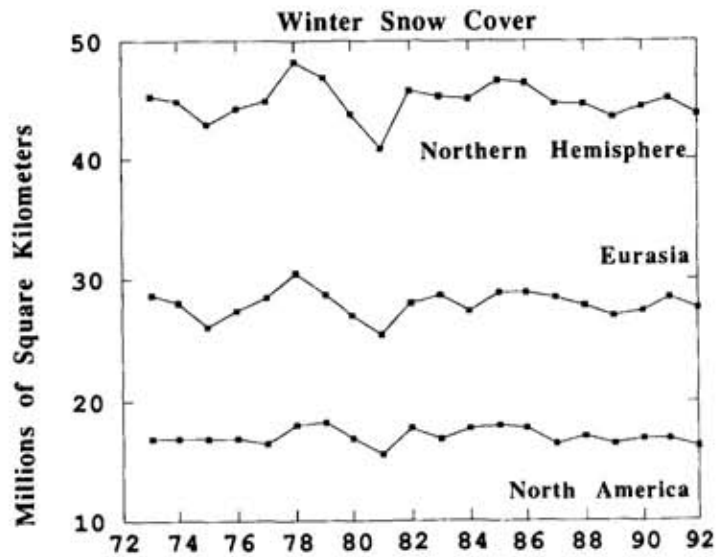
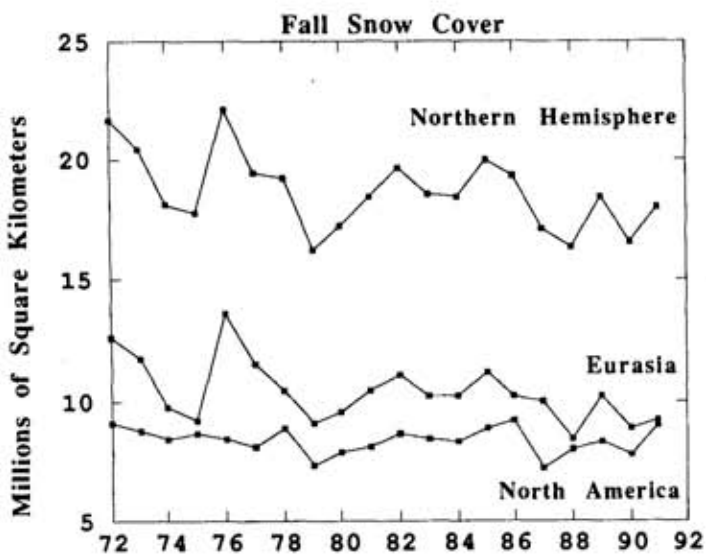
Twelve-month running means of continental snow extent best illustrate the periods of above normal cover that occurred in the late 1970s and mid 1980s (fig. 7). Intervals with lower snow extents include the mid 1970s and early 1980s, however neither approach the deficit of snow cover observed in recent years. Of the 58 months between August 1987 and May 1992, only five had above normal snow cover (Jan 88, Sep 89, Dec 89, Dec 90, Nov 91). The lowest year on record was 1990, when monthly minima occurred in eight months (table 1). Spring cover has shown pronounced deficits over the past five years in Eurasia and six years in North America; areas in these Springs have been at or below lows established prior to this period (fig. 8). During the same interval, both continents have had low seasonal cover in the Fall and Summer, although frequently neither continent has been at or approached record low levels. Winter cover has been close to average over the past six years.



**Figure 6.** Monthly snow cover over northern hemisphere lands (including Greenland) between January 1972 and May 1992. The median area of cover is the horizontal line within the twelve monthly boxes, and the interquartile range (ICR) is between the top and bottom of the box. Whiskers show the extreme values between  $\pm 1$  and  $\pm 1.5 * ICR$ , and asterisks show values outside this range. Values are calculated from NOAA weekly snow charts using the Rutgers Routine.



**Figure 7.** Twelve-month running means of snow cover over northern hemisphere lands (including Greenland) for the period January 1972 through May 1992. Running means are also shown for Eurasia and North America (including Greenland). Values are plotted on the 7th month of the 12 month interval, and are calculated from NOAA weekly snow charts using the Rutgers Routine.



**Figure 8.** Seasonal time series of snow cover over Eurasia and North America (Greenland is excluded). Values are calculated from NOAA weekly snow charts using the Rutgers Routine.

**Microwave Charting**

Microwave radiation emitted by the earth's surface penetrates winter clouds, permitting an unobstructed signal from the earth's surface to reach a satellite. The discrimination of snow cover from microwave data is possible mainly because of differences in emissivity between snow-covered and snow-free surfaces. Estimates of the spatial extent as well as



the depth or water equivalent of the snowpack are gleaned from equations employing radiation sensed by multiple channels in the microwave portion of the spectrum (e.g., Kunzi et al., 1982; McFarland et al., 1987). Snow estimates from satellite-borne microwave sensor data have been available since the launch of the Scanning Multichannel Microwave Radiometer (SMMR) in late 1978. The spatial resolution of the data is on the order of several tens of kilometers. Since 1987, close to the time of SMMR failure, the Special Sensor Microwave Imager (SSM/I) has provided information for the determination of snow extent and volume. The lack of sufficient ground truth data on snow depth or volume makes an adequate assessment of the reliability of such microwave estimates uncertain. Therefore the remainder of this discussion focuses on the microwave monitoring of snow extent.

As with shortwave products, the microwave charting of snow extent is not without its limitations. The resolution of the data makes the detailed recognition of snow cover difficult, particularly where snow is patchy, and it is difficult to identify shallow or wet snow using microwaves. It is also apparent that because of region-specific differences in land cover and snowpack properties, no single algorithm can adequately estimate snow cover across northern hemisphere lands.

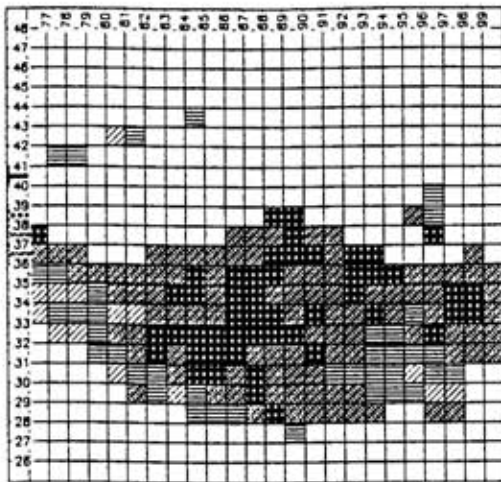
### ***Regional Investigations***

Efforts are underway to better understand regional microwave signatures, and in some cases to develop region-specific algorithms. Landscapes of interest include mountains (Chang et al., 1991; R. Armstrong, per. commun.), forest (Hall et al., 1982; Hallikainen & Jolma, 1986), tundra (Hall et al., 1986) and prairie (Goodison, 1989). Also, over the Tibetan Plateau and adjacent high mountains of south central Asia, snow cover tends to be overestimated when a single hemispheric (generic) algorithm is employed (Robinson et al., 1984). In a recent investigation, the generic algorithm of Chang et al. (1987) (hereafter, NASA) is adjusted using shortwave-derived charts and GIS techniques to better represent snow conditions in this region (Robinson et al., 1990).

The theoretical NASA algorithm uses the difference in brightness temperatures of 18 and 37 GHz SMMR data to derive a snow depth/brightness temperature relationship for a uniform snow field. A snow density of 0.3 g/cm<sup>3</sup> and a snow crystal radius of 0.3 mm are assumed, and by fitting the differences to the linear portion of the 18 and 37 GHz responses a constant is derived that is applied to the measured differences. This algorithm can be used for snow up to one meter deep. The shortwave charts are constructed expressly for this study, and for the purpose of this study are considered accurate. Microwave and shortwave data are reduced to 1° x 1° cells for this analysis.

Given the linear nature of the NASA algorithm, snow cover is subtracted from all cells in one centimeter increments until the extent of snow cover in the microwave chart is in close spatial agreement with the shortwave product. In 15 Plateau cases, between 4 and 16 cm of snow have to be subtracted (offset) from original cell totals to reach closest agreement with the shortwave chart. The mean microwave offset for the Plateau is 8 cm, while for the adjacent high mountains the mean is 6 cm, with cases ranging from 0 to 16 cm. Figure 9 shows the shortwave, generic microwave and adjusted (to the mean offset) microwave charts for the two regions for January 1987. The 8 and 6 cm offsets compare to mean offsets of 0 to 4 cm in the lower elevation basins and lower mountains elsewhere in western China.

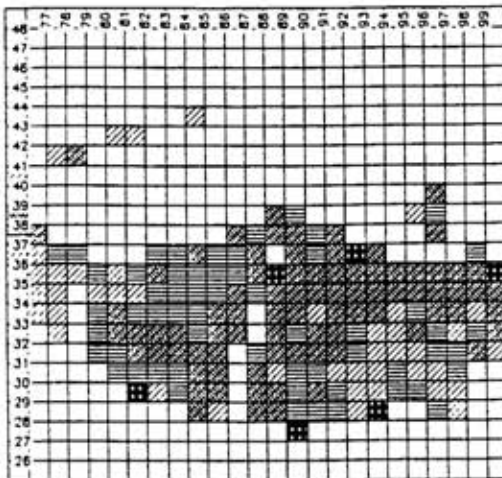
A



A) The shortwave chart is constructed using Defense Meteorological Satellite Program imagery and shows 1° x 1° cells with, 1) (cf. legend) patchy (10-30% cover), 2) partial (30-80%) and 3) full (80-100%) snow cover, along with 4) areas of snow-free ground (<10% snow cover). Empty cells are outside the study region.

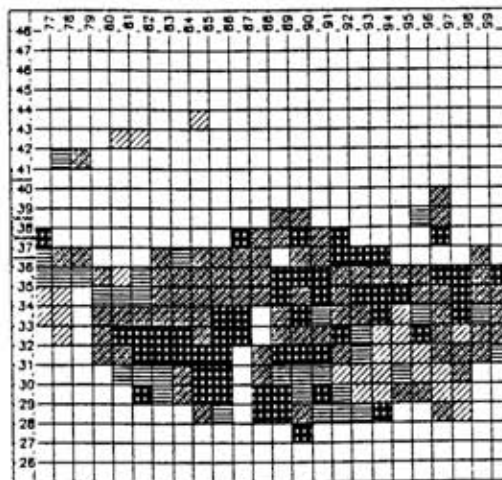


B



B) The generic microwave chart is constructed using the algorithm of Chang et al. (1987) and shows cells with snow depths ranging from, 1) 3-10 cm, 2) 11-20 cm and 3) >20 cm, along with 253) snow-free (<3 cm snow) ground. Empty cells are outside the study region or where data within the region is missing.

C



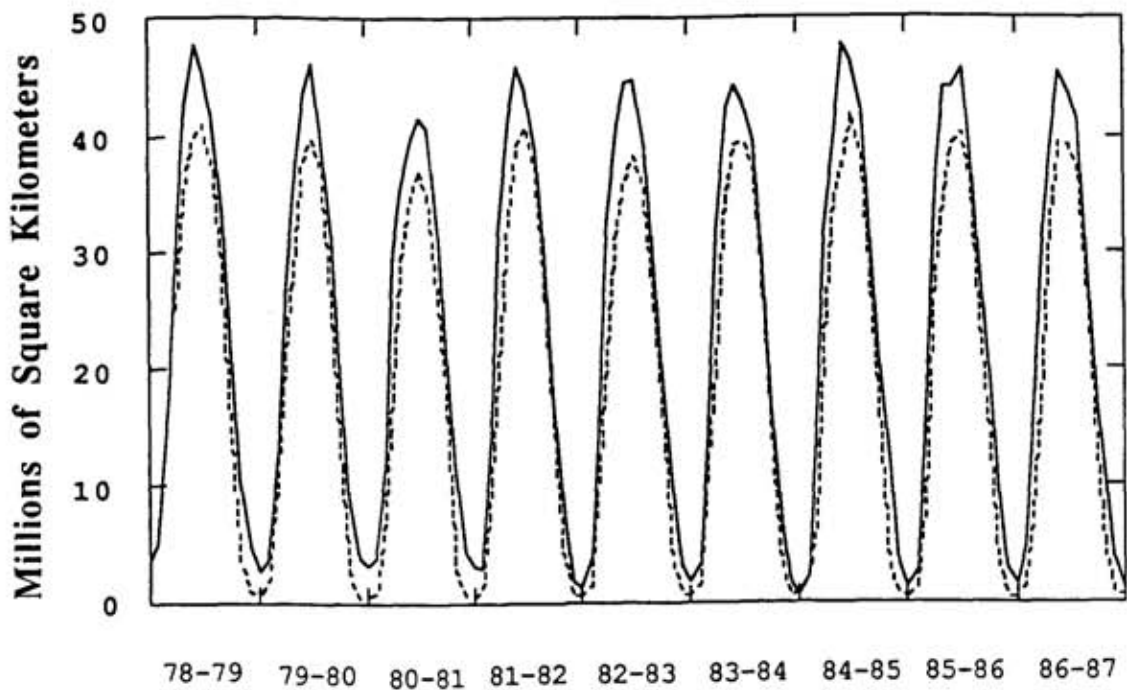
C) The adjusted microwave chart shows the cells considered to be snow covered once 8 cm of snow is subtracted from the generic reports over the Plateau and 6 cm subtracted over the high mountains. Cover categories are the same as for the generic chart.

**Figure 9.** The shortwave, generic microwave and adjusted (to the mean offset) microwave charts for the Tibetan Plateau and surrounding high mountains for mid-January 1987.

The overestimation of snow cover when employing the generic NASA algorithm over this high elevation area may be associated with the patchiness and general shallow depth of the snow cover in the area, or to particular characteristics of the snowfree surfaces in the region. However, it appears that the major cause of the overestimation has to do with the thin atmosphere at these elevations. The generic algorithm was developed and tested for low elevation sites, and by removing the atmosphere from the theoretical algorithm, the match between the microwave and shortwave products is much improved (A. Chang, per. commun.). However, this comes at the expense of adjacent lower elevation regions, where the revised microwave snow area estimates are low.

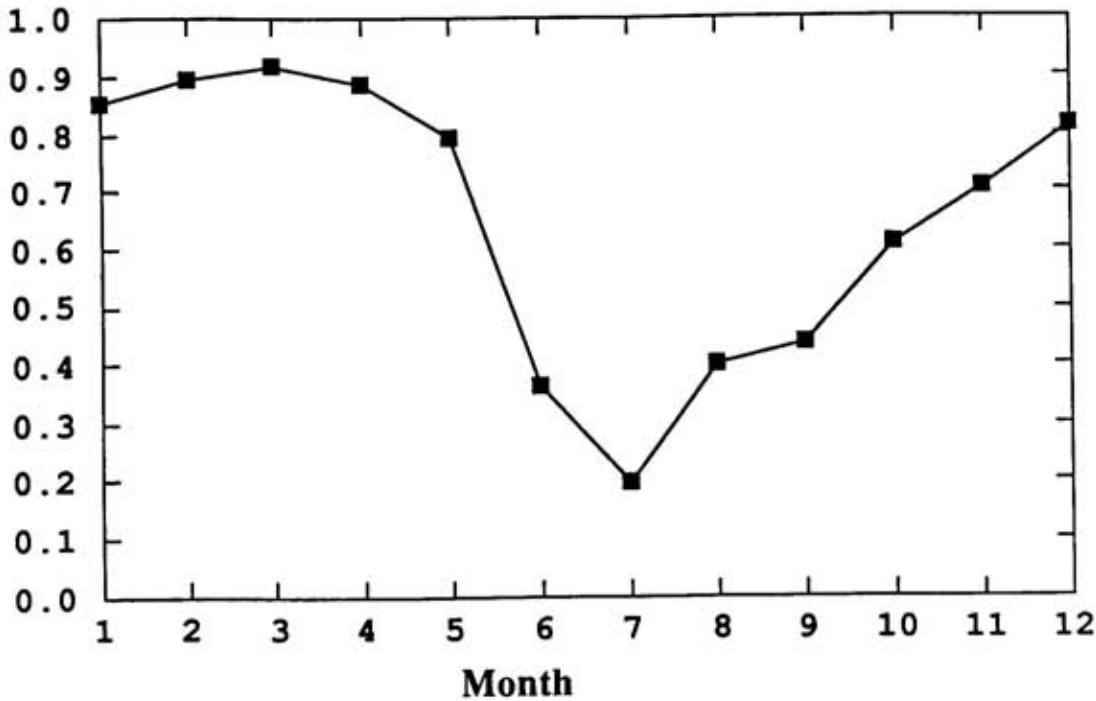
***Continental snow cover from microwave charts: 1978-1987***

The difficulties in monitoring snow extent over Tibet using the generic NASA algorithm are rather unique in that they show an overestimation of snow. Most of the errors in microwave sensing are in the opposite direction. This is apparent when comparing microwave estimates of monthly continental snow extent with NOAA shortwave values (Chang et al., 1990). This is the only such time series available to date, and covers the interval from November 1978 through August 1987. NASA mean monthly snow cover for northern hemisphere lands (exclusive of Greenland) runs from less than one to as much as thirteen million square kilometers below NOAA areas for the nine years of coincidental estimates (fig. 10). These absolute differences are greatest in the late Fall and early Winter.



**Figure 10.** Extent of snow cover over northern hemisphere lands (excluding Greenland) computed from NOAA shortwave charts using the Rutgers Routine (solid line: Sep 78-Aug 87) and from NASA microwave charts (dashed: Nov 78-Aug 87).

In a relative sense, microwave areas are between 80 and 90% of shortwave values in Winter and Spring, 20 to 40% of the shortwave estimates in Summer, and 40 to 70% of shortwave areas in Fall (fig. 11). A possible explanation for the great disparities in the latter two seasons may be the wet and shallow nature of the snowpack interfering with accurate microwave recognition of snow. Depth may be the most important of the two variables, given the better agreement in Spring, although it has been suggested that unfrozen soil beneath the pack is a major contributor to the underestimates during Fall (B. Goodison, per. commun.)



**Figure 11.** Fractional difference between monthly snow cover area derived from NOAA shortwave and NASA microwave charts for northern hemisphere lands (excluding Greenland). NOAA values, derived using the Rutgers Routine, exceed NASA estimates for each month within the November 1978 through August 1987 study period.

While the preceding discussion has dealt with SMMR data, snow monitoring employing SSM/I data shows the same strengths and liabilities as the former (Goodison, 1989; Hall et al., 1991). The SSM/I has 19 and 37 GHz channels, thus SMMR algorithms perform much the same as they do with the 18 and 37 GHz channels. In addition, the 85 GHz channel on the SSM/I has shown promise in improving the monitoring of shallow (<5 cm) snow cover (Nagler & Rott, 1991).

### Snow Cover Over Arctic Sea Ice

Most studies of snow on top of arctic sea ice have been of limited spatial and temporal scope. They have primarily been concerned with the snow melt season and the resultant impact of snow melt on the surface albedo of the pack ice. Studies have been based largely

on observations at drifting stations, on fast ice, and during aircraft missions (e.g., Kuznetsov & Timerev, 1973; Pautzke & Hornof, 1978; Hanson, 1980). Others have used these measurements (e.g., Marshunova & Chernigovskiy, 1966; Hummel & Reck, 1979; Kukla & Robinson, 1980) and satellite passive microwave data (Carsey, 1985) to estimate regional summer albedos. These albedos, to a large extent, are dictated by the condition and distribution of the snow cover on top of the ice, particularly in the Spring and early Summer.

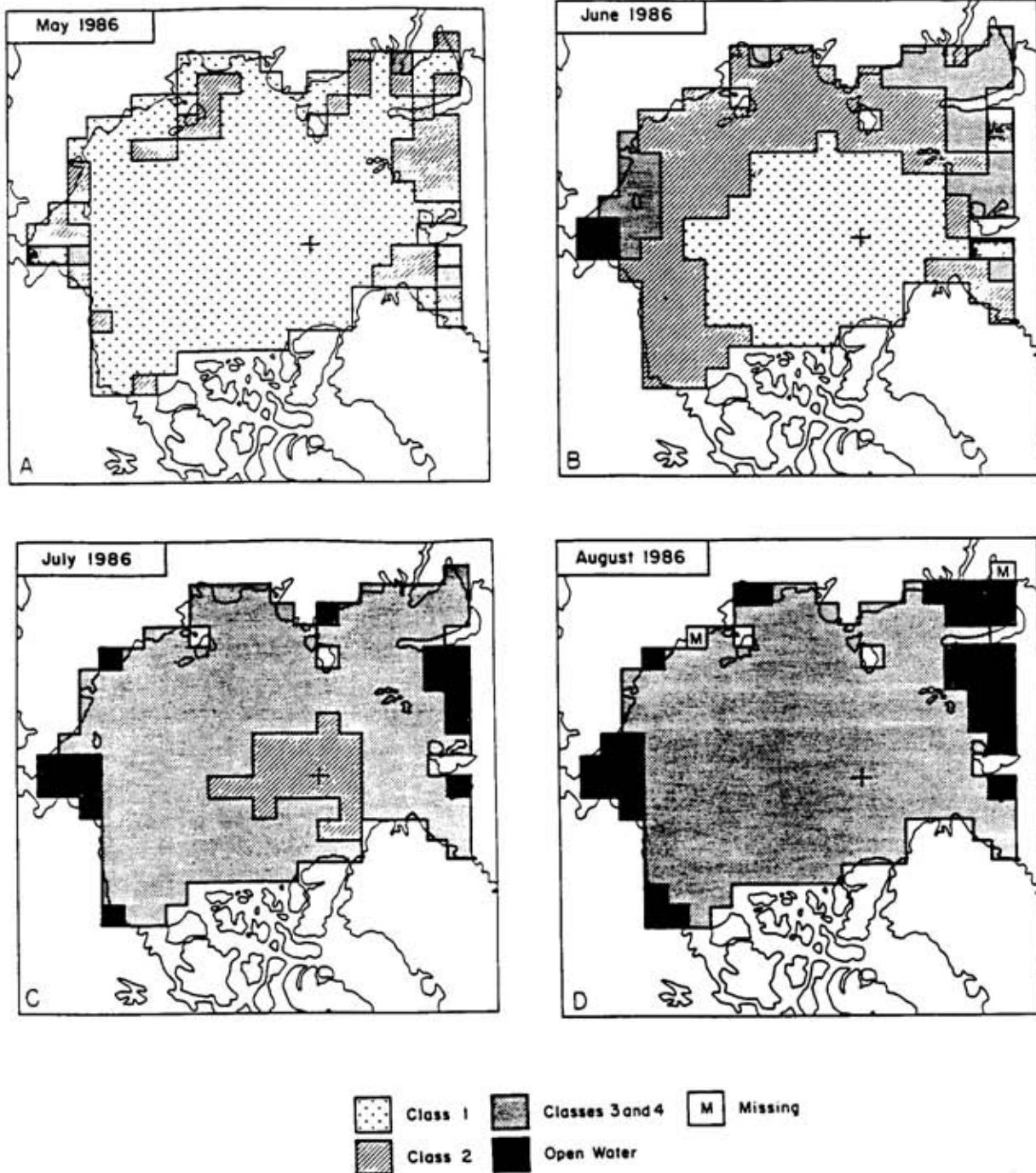
Only one time series of snow conditions on top of arctic sea ice has been produced to date (Robinson et al., 1992). This effort employs shortwave satellite imagery to map manually surface brightness changes over sea ice throughout the Arctic Basin from May to mid-August over a ten year period. Due to gaps in available imagery, the ten years are not continuous, falling between 1975 and 1988. Only melt is mapped, as cloudiness and low solar illumination preclude charting of the onset of fall snow cover across the basin.

The year 1986 typifies the pattern of snow melt found in most years (fig. 12). Snow melt is observed to begin in May in the marginal seas, progress northward with time, and finally begins near the Pole in late June. In most years, snow melt is completed in the central arctic by late July. Large year-to-year differences occur in the timing of snow melt, exceeding one month in some regions. The onset of melt, defined by the date when at least 50% of all non-water cells in a given portion of the basin are class 2 or greater (cf. fig. 12 caption for class descriptions), is observed to vary from mid May in the Kara and Barents seas to late June in the central Arctic (table 2, fig. 13). Melt onset was particularly early in 1977 and 1984 and late in 1975 and 1979. The onset of advanced melt (>50% of a region class 3 or greater) starts about one month after melt onset in all but the central arctic, where it is three weeks later, and in the East Siberian/Laptev area, where it is reached within one week of melt onset. Within a region, the year-to-year variation in the onset of advanced melt is from approximately two to four weeks, somewhat less than variations in melt onset. For most of the basin, advanced melt was earliest in 1987, followed closely by 1977, and latest in 1978.

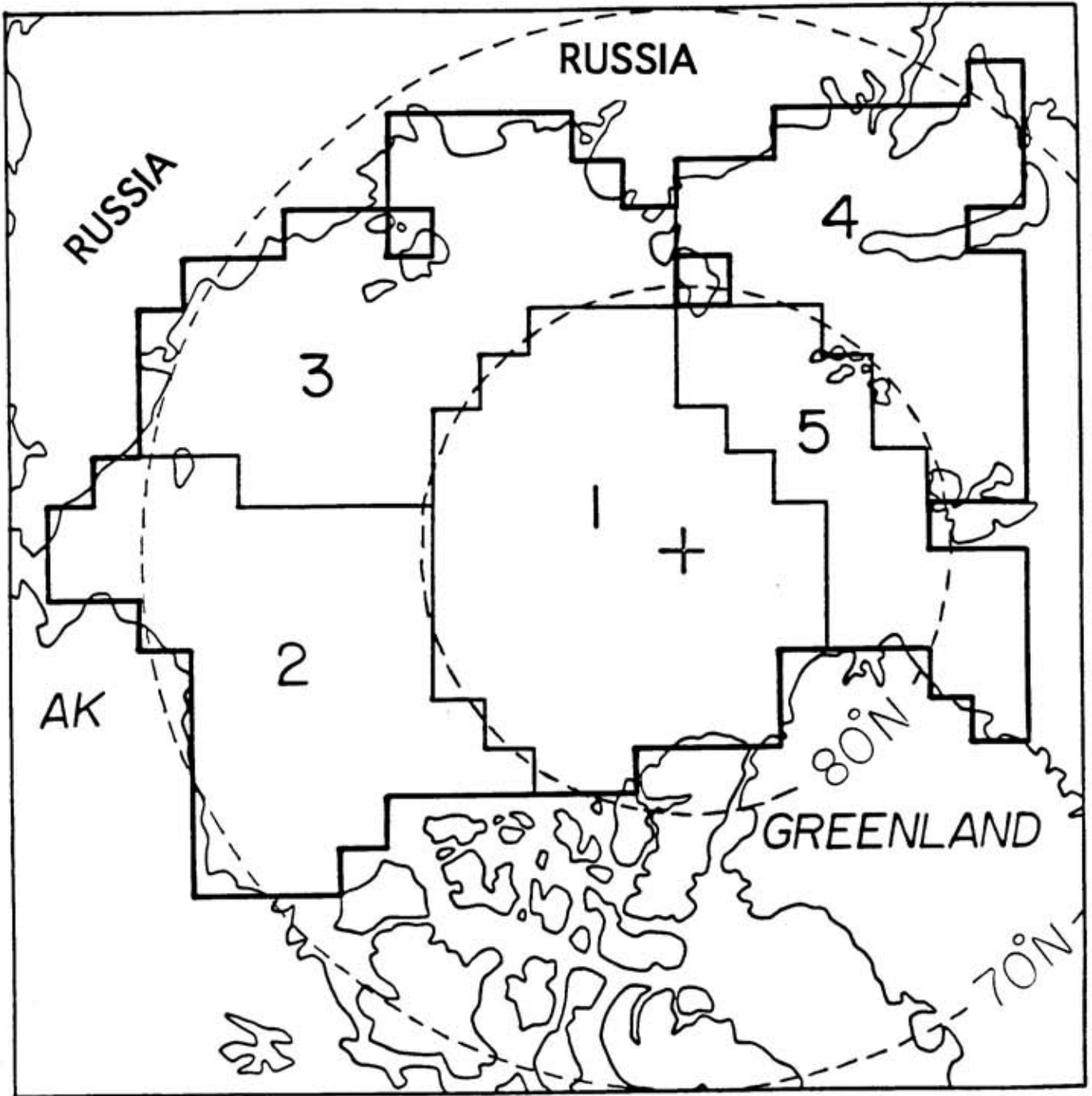
**Table 2.** Dates of the onset of spring snow melt on top of sea ice in five regions of the Arctic Basin (cf. fig. 13). Mean dates and the range in dates (days) between extreme years are also given. Dashes denote those years where cloudiness precludes an accurate estimate of onset.

	<b>Central Arctic</b>	<b>Beaufort Chukchi</b>	<b>Laptev E. Siberian</b>	<b>Kara Barents</b>	<b>NW North Atlantic</b>
1975	7/8	6/14	6/29	5/12	6/8
1977	6/5	5/21	5/21	5/12	5/12
1978	6/23	5/24	6/8	5/18	5/24
1979	6/29	6/20	6/17	5/30	6/2
1980	7/2	6/11	6/23	5/15	6/23
1984	6/14	5/9	6/11	5/3	5/6
1985	6/20	6/8	6/14	5/27	5/27
1986	7/5	6/11	6/14	5/6	-
1987	6/11	5/30	6/5	5/12	5/18
1988	6/14	5/9	6/17	-	-
Mean	6/23	5/30	6/14	5/15	5/27
Range	27 d	42 d	39 d	24 d	48 d





**Figure 12.** Progress of arctic snow melt during 1986 as shown by the mean brightness classes for study grid cells in the basin for, A) May, B) June, C) July, and D) 1-16 August. Class 1 corresponds to fresh snow cover over 95% of the ice and class 2 is found when snow covers between 50-95% of the surface, with the remainder being bare or ponded ice. Class 3 represents the advanced to final stage of snow melt, with numerous meltponds and between 10-50% of the ice surface snow covered, or, following pond drainage, predominantly bare ice. Heavily-ponded or flooded ice is represented by class 4, and is generally limited to regions of fast ice near outlets of major rivers along the Siberian coast.



**Figure 13.** Arctic Basin study area, including, 1) Central Arctic, 2) Beaufort/Chukchi seas, 3) East Siberian/Laptev seas, 4) Kara/Barents seas, and 5) NW North Atlantic subregions.

Microwave remote sensing over arctic sea ice has shown promise in identifying what may be the initial appearance of liquid water within the snow pack or at the snow/ice interface (Anderson & Robinson, 1985; Cavalieri et al., 1990). This signal is recognized approximately three weeks prior to the onset of melt as observed using shortwave imagery. Further study is needed to better understand this signal and to integrate this information with shortwave data.

## **Snow Monitoring In The 1990s**

Much remains to be accomplished in the 1990s to assure that accurate information on the extent and other characteristics of snow cover is available on regional and hemispheric levels. This includes ongoing efforts to monitor surface conditions during the decade and beyond, as well as retrospective efforts to improve estimates of past snow covers. The following discussion is not meant to be exhaustive, rather it highlights some of the more important work across a broad spectrum of needs.

### ***Ongoing Monitoring***

- 1) Maintain the NOAA shortwave satellite charting effort in its present form. To abruptly or, perhaps more seriously, subtly alter the manner in which these charts are produced would severely weaken what is presently the longest and most consistent satellite-derived data set of any surface or atmospheric variable. Any alteration would require "restarting the clock" and losing precious years when employing snow cover as a means of identifying and monitoring climate change.
- 2) Produce all-weather, all-surface snow charts using shortwave and microwave satellite data and station observations. The maintenance of the shortwave effort need not exclude the development of new and no doubt more accurate means of monitoring snow extent. For instance, geographic information systems techniques should be developed to produce a new series of operational hemispheric snow charts on at least a weekly basis. These charts would provide the most detailed information possible on snow extent, water equivalent and depth, and surface albedo over land surfaces, and, at the least, snow extent and albedo over sea ice and the Greenland ice sheet.

To begin with, these charts need not employ a consistent methodology. Rather they should be designed to accommodate improvements in all the realms of snow monitoring, be they increased station coverage, operational satellite monitoring of the surface at 1.6  $\mu\text{m}$ , new regional microwave algorithms, or improved GIS techniques. The current NOAA product can handle the time series duties over the continents for the time being, and time series of snow over sea ice should be continued using proven techniques.

### **Retrospective Monitoring**

- 1) Continue the assembly, digitization and quality control of historic to recent station data from throughout middle and high latitude lands. Data from snow courses and remote snow measurement networks (i.e., mountain observations) should also undergo the same scrutiny.
- 2) Construct regional snow charts from ground data. Interpolative techniques, which account for horizontal and vertical variations, should be used to estimate snow extent and depth for regions with sufficient data. This charting should extend back as far as possible and continue into the microwave satellite era. Charts within the satellite era



would permit cross checks of extent and depth estimates with shortwave and microwave products.

- 3) Utilizing data from urban and rural locations, study the impact of urbanization on snow cover depth and duration.
- 4) Rechart continental snow cover for the 1966 to 1971 interval. Recharting is also recommended for the Tibetan Plateau and adjacent mountains through the middle 1970s and for high-latitude lands during the Summer. These endeavors, along with spot checks across the continents in subsequent years, would expand and improve the NOAA time series and would permit the establishment of quantitative error ranges to these charts.
- 5) Produce a time series of surface albedo over northern hemisphere lands. Using GIS techniques, NOAA weekly snow chart data should be merged with available information on the surface albedo of snow covered and snowfree landscapes.
- 6) Once greater confidence in regional microwave algorithms is attained, a reassessment of hemispheric snow cover using SMMR and SSM/I data should be undertaken. This should include estimates of snow extent and depth and/or water equivalent.
- 7) Expand the ten-year data set of snow extent over arctic sea ice to include as many years as possible between the early 1970s and present.

## **Conclusions**

The critical role that snow cover plays in the global heat budget and the expected role of snow feedbacks in anthropogenically-induced climate change support the continued diligent monitoring of snow cover over continents and sea ice. With the availability and better understanding of data from a variety of satellite and ground sources, and the ability to integrate and examine these data using geographic information techniques, more accurate and extensive knowledge of snow cover across the Northern Hemisphere is within reach. This information, along with expanded retrospective analysis efforts, will solidify the position of snow as one of the key indicators of future change in the climate system.

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Walsh, J.E., W.H. Jaspersen & B. Ross. 1985. Influences of snow cover and soil moisture on monthly air temperature. *Mon. Wea. Rev.*, 113, 756-768.



# Northern Hemisphere Snow Cover and Temperature Patterns in the 1980's

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## Introduction

Estimates of global surface temperature based on meteorological data indicate that the decade of the 1980's has been the warmest such period in the instrumental record. Much less attention has been focused on the patterns of the temperature anomalies during this period even though the most simple analyses readily show that the temperature anomalies are not uniform in magnitude or sign over the globe, Ropelewski et al, 1992. In this paper we provide a description of the patterns of global temperature anomalies during 1980's and present some preliminary analyses to examine the relationship between these anomaly patterns and decadal mean snow cover anomalies.

## Data

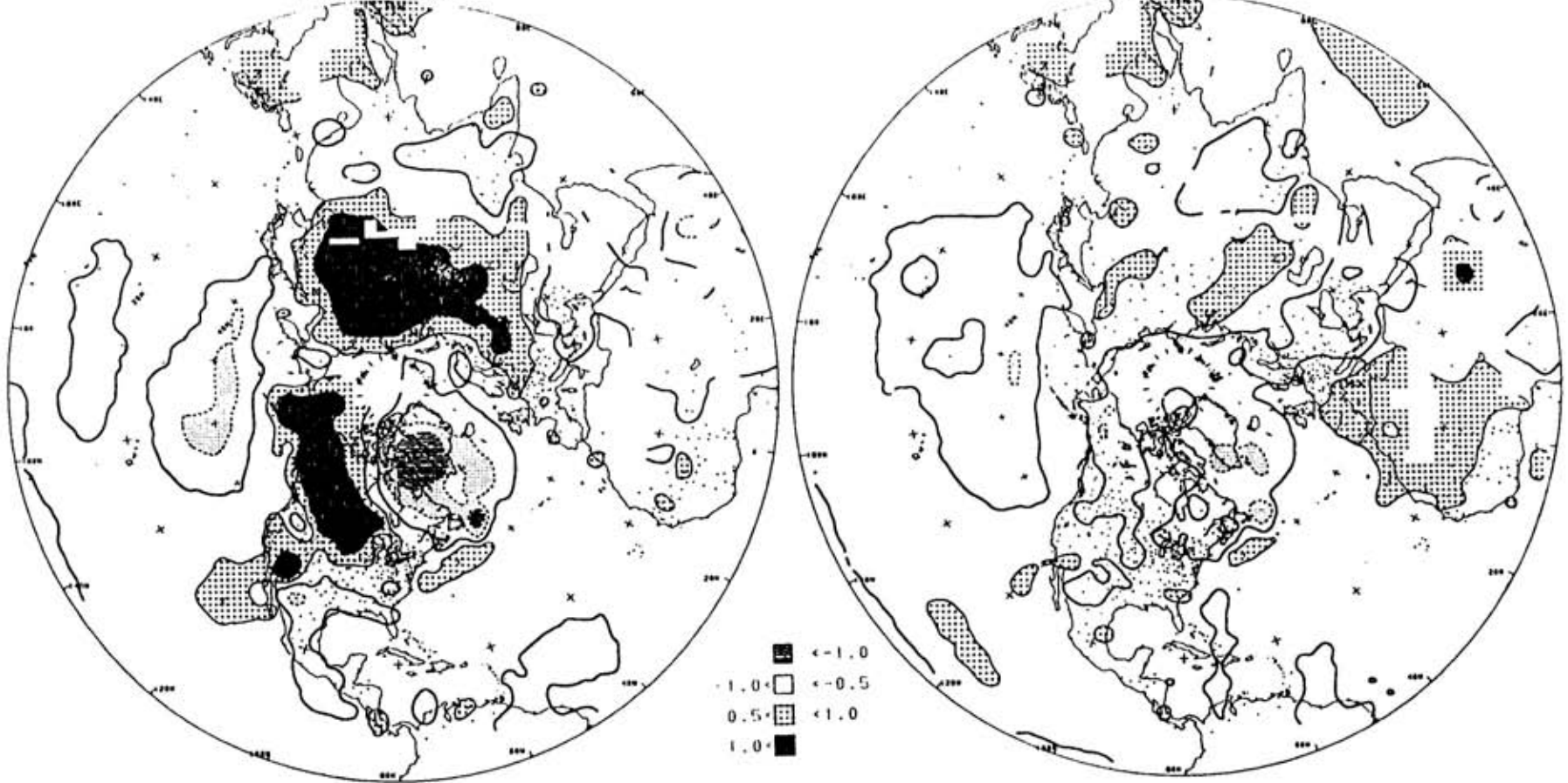
Monthly surface temperature data for the period 1950 to 1990 were obtained from the Climate Anomaly Monitoring System (CAMS) data base, Ropelewski et al., 1985. Sea surface temperatures are taken from the NMC/CAC operational analyses and compared to the COADS/ICE climatology, Reynolds, 1988. Monthly snow cover data for the period 1973 to 1990 were derived from weekly operational analyses of satellite snow cover maps, Ropelewski, 1991. Monthly snow cover anomalies were recomputed from weekly archived values using a consistent set of land-sea masks and a consistent algorithm. While these corrections differ slightly from those discussed by Robinson (this volume) these differences are small and should not substantially alter the analyses.

## Decadal Temperature Anomaly Patterns

Analyses of global and hemispheric temperature anomaly time series indicate that the largest anomalies tend to occur in the Northern Hemisphere winter and spring and, further, that the global signal is dominated by the Northern Hemisphere, Jones, (1988), Halpert and Ropelewski, (1991). For the 1980's, large differences are evident in the decadal temperature anomaly patterns between the first and second half of the year, Fig. 1. The December - May period shows temperature anomalies greater than 1°C covering Alaska and western Canada and another large area with positive anomalies over Russia. Strong negative anomalies were found over Greenland. In contrast, during the June - November season, small negative anomalies were found over Alaska and western Canada. These mean patterns suggest a reduction in the amplitude of the annual cycle over northwestern North America during the past decade. Negative SST anomalies in the North Pacific and positive temperature anomalies in North America, for the December to May season, have been statistically associated with ENSO, Ropelewski and Halpert, (1986), Halpert and Ropelewski (1992). However, preliminary examination of the December to May temperature anomalies for individual years during the 1980's indicates that the large positive temperature anomalies over northwestern North America were also present during the non-ENSO years of the 1980's. Thus, the pattern of decadal mean temperature

Dec - May

Jun - Nov



**Figure 1.** Decadal (1981-1990) mean surface temperature anomaly patterns for the December through May season (left) and June through November season (right). The analysis is based on station data over land and sea surface temperature over water. Anomalies for station data are from the 1951-1980 base period, and from the COADS/ICE climatology (Reynolds, 1988) over water. Small plus signs indicate data locations over land. The contour interval is  $0.5^{\circ}\text{C}$  with negative anomalies dashed.

anomalies over North America can't be ascribed to ENSO. In the Northern Hemisphere warm season (June to November) the decadal temperature anomaly patterns were less pronounced with the largest area of warm anomaly over Northern Africa and Southern Europe. It is likely that a part of the African temperature anomaly pattern is a reflection of the decadal scale drought in the Sahel. The magnitudes of decadal SST anomalies were generally small and positive, except in an area of the North Pacific and in the North Atlantic around Greenland.

### **Snow Cover Anomaly Patterns**

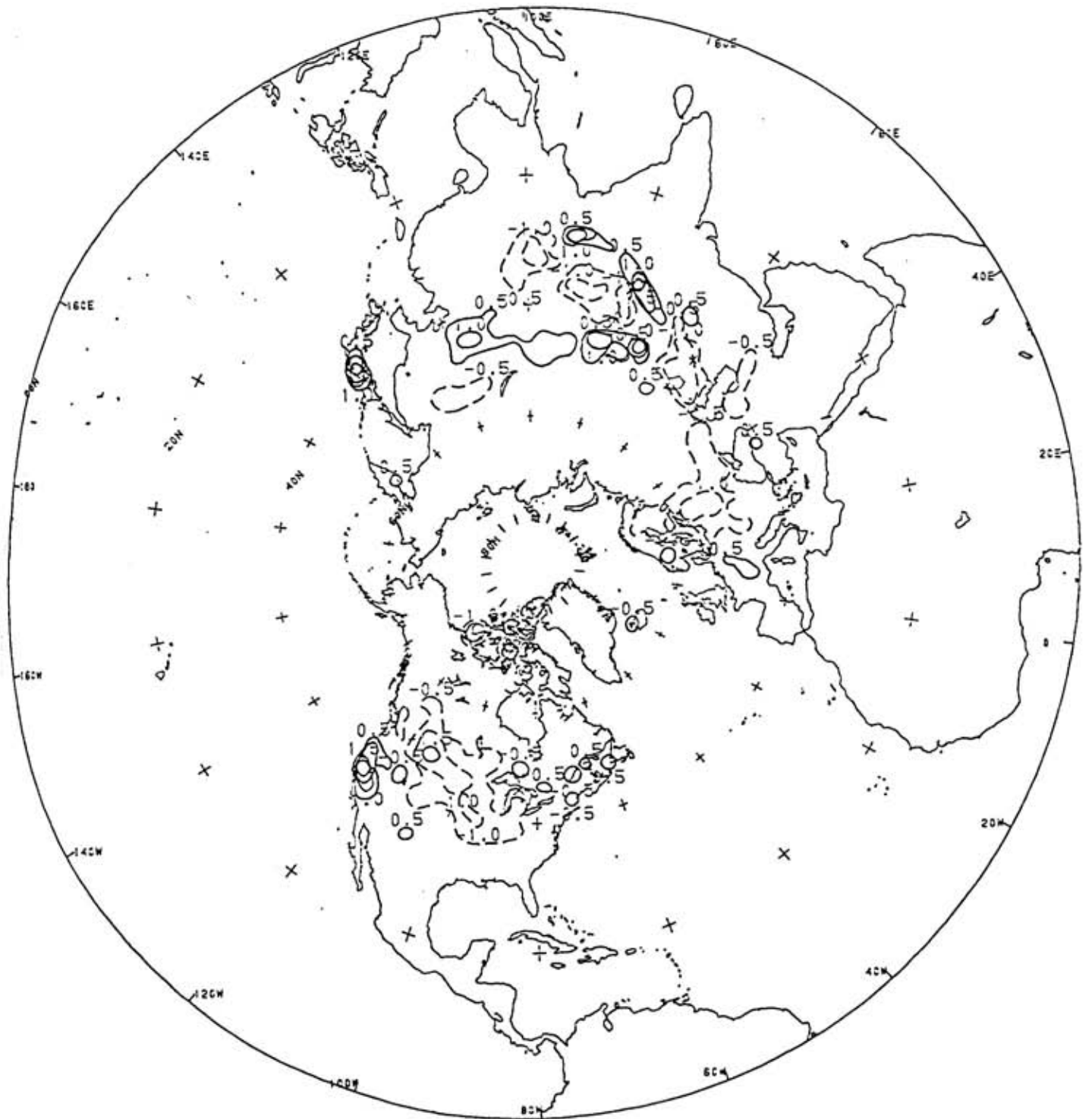
March 1990 had the largest positive monthly anomaly in the global temperature time series and contributed significantly to 1990 being the warmest year on record, e.g. Halpert and Ropelewski, (1991). The largest negative snow cover area anomaly in the 25 year satellite record also occurred in March 1990, suggesting a relationship between global temperatures and snow cover, Robinson et al., 1991. Preliminary inspection revealed that the implied snow cover temperature relationships were strongest over Eurasia in the spring, e.g., Fig. 2. A comparison of the mean temperature anomaly patterns (Fig. 1) and snow cover anomaly patterns (Fig. 3) shows relatively good correspondence between snow and temperature anomalies over North America but a more complicated picture for Eurasia. Closer inspection, however, reveals that the largest negative North American snow cover deficits fall on the southern flanks of the largest positive temperature anomalies. This suggests that, over western Canada and Alaska the positive temperature anomalies of the 1980's were not associated with corresponding deficits in snow cover. Of course, these can't address the questions regarding snow depth or liquid water equivalent. The snow-temperature relationships are even more complex in the Eurasian sector. Even though the time series, Fig 2., suggest a simple inverse relationship between snow cover and temperature, the snow anomaly charts indicate that some of the Eurasia areas with the largest positive temperature anomalies also experienced positive snow cover area anomalies. Also, the largest areas of snow cover deficit during the 1980's lay in the European sector, or to the west, of the largest positive temperature anomalies.

### **Discussion**

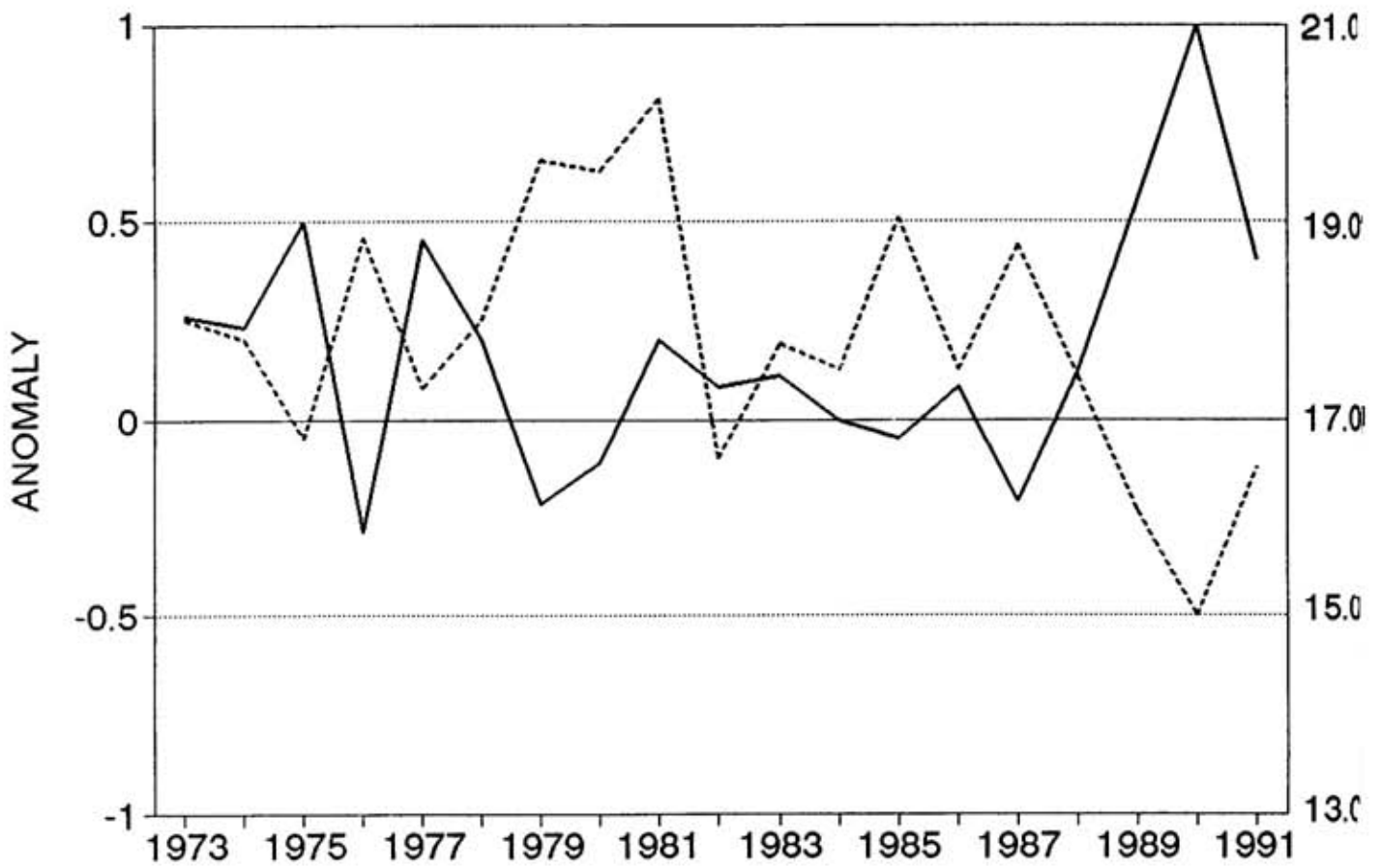
The preliminary analysis outlined here suggests the Northern Hemisphere temperature anomalies during the 1980's are dominated by first half of the year, i.e., December - May. The preliminary analysis suggests that Northern Hemispheric surface temperature anomalies may be related to snow cover anomalies but that this relationship may not be simple. On the global and continental scale, anomalies of temperature and snow cover area tend to be anti-correlated. However, over the decade of the 1980's, the largest areas of snow cover anomalies did not coincide, in general, with the areas covered by the largest temperature anomalies. A more complete analysis is needed to quantify and understand the relationships between snow cover anomaly and hemispheric and global snow cover.

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**Figure 2.** Northern Hemisphere December-to-May, i.e., cold season, mean snow cover anomaly for 1981-90 with respect to the 1973 to 1990 base period. Dashed (solid) outline areas experienced below (above) normal snow cover.



**Figure 3.** Time series of a) dashed line - Eurasian snow cover area (105 km<sup>2</sup>) derived from analysis of satellite imagery (Matson et al., 1986) and b) solid line - Eurasian temperature anomaly index, both for the spring.

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# Temperature Variations in Northern High Latitudes, 1960-1990

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## Abstract

Feedbacks resulting from the retreat of snow cover and sea ice contribute to the polar amplification of the greenhouse warming projected by global climate models. Low-frequency variations or trends of surface air temperature can trigger these feedbacks. High-latitude air temperatures for the Northern Hemisphere are examined here for indications of changes over the past three decades, 1960-1990. The data show a distinct warming over northern land areas during winter and spring, but little or no warming over the subpolar oceans. The pattern is not inconsistent with results of the more recent greenhouse experiments performed with coupled atmosphere-ocean models. The changes of air temperature are spatially compatible with changes of sea ice over the same 30-year period. The changes of total hemispheric sea ice extent since 1960 show a decrease in the spring and summer seasons. The largest decrease of ice extent has occurred in the past few years.

## Background

A pervasive feature of global climate model simulations with enhanced greenhouse (e.g., CO<sub>2</sub>) forcing is an amplified warming in the polar regions, especially near the surface and in the winter half of the year. Atmospheric models that are not coupled to a deep ocean project wintertime warmings of 8-16°C in portions of the Arctic and Antarctic when the CO<sub>2</sub> concentration is doubled (IPCC, 1990, Plate 5.4). This rather dramatic warming, which is stronger than that projected for other parts of the hemisphere, is at least partly attributable to the temperature-albedo feedback associated with the retreat of snow and sea ice. The simulated warming is indeed generally largest in the areas from which snow and ice retreat. Other factors contributing to the polar amplification are the breakdown of the shallow but steep near-surface temperature inversions in the polar regions and the trapping of terrestrial radiation by the increased water vapor and cloudiness that accompany the warming.

Coupled models containing interactive (deep) ocean components also produce the strongest warming in high latitudes during winter, although the model-to-model differences of the projections can be large (IPCC, 1990, p. 182-183). There is also a tendency in at least some coupled models for a reduced warming over the North Atlantic subpolar seas and over the Southern Ocean. Factors tending to reduce the surface warming over the subpolar oceans relative to that over land include the ocean's thermal inertia, a tendency for sea ice to form more readily because subpolar surface waters are freshened by increased precipitation (Manabe et al., 1991), and changes in the subpolar fields of sea level pressure and surface wind stress (Washington and Meehl, 1989). The situation in the subpolar North Atlantic is further complicated by the tendency for at least some models to possess two stable equilibrium states, one with a vigorous thermohaline circulation in the North Atlantic and the other with little or no thermohaline circulation (Manabe and Stouffer, 1988).

The areal coverage of continental snow cover decreases substantially in these simulations, especially in the winter and spring seasons, although the magnitude (and even the sign) of the albedo-temperature feedback varies considerably from model to model (Cess et al., 1991). The models also project a substantial decrease of sea ice. In some simulations with doubled CO<sub>2</sub>, the sea ice in the Arctic disappears almost completely during the summer (e.g., Manabe et al., 1992; Wilson and Mitchell, 1987).

Despite differences in the longitudinal dependences, all models indicate at least some polar amplification of the greenhouse warming. Thus the variation of polar climate over the past several decades assumes particular importance in the context of the detection of possible greenhouse effects. In this paper, we attempt to complement the Workshop's emphasis on snow cover by summarizing several decades of recent data on the air temperature and sea ice cover of the northern high latitudes. Recent air temperature variations have been discussed elsewhere (e.g., IPCC, 1990; Hansen and Lebedeff, 1987; Jones et al., 1986), but with no emphasis on the polar regions. (Spatial maps of temperature variations and their temporal changes are usually displayed on Mercator map projections, for example.) Decadal-scale trends in sea ice have received little attention, largely because sea ice datasets span shorter periods than do temperature datasets. As will be apparent in the subsequent sections, however, existing sea ice datasets are adequate to depict recent variations that are generally consistent with those of air temperature and snow cover.

## Data and Derived Variations

### *Surface air temperature*

Recent variations of high-latitude temperatures have been evaluated from the dataset produced by the Climate Research Unit of the University of East Anglia (Jones et al., 1986). The Climate Research Unit produced this dataset by consolidating (1) monthly air temperatures from land surface stations and (2) monthly sea surface temperatures from the Comprehensive Ocean-Atmosphere Dataset. The latter dataset, which is based on ship reports, is described by Woodruff et al. (1987). Jones et al. have examined the station data for contamination by site relocation, urbanization and other changes; eliminated suspect data; and consolidated the land and ocean data into 5° X 5° latitude-longitude cells.

We have used this dataset to evaluate best-fit linear trends over the 1960-1990 period for all grid cells (poleward of 40°N) for which no more than 20% of the monthly values were missing. These trends were evaluated for both the annual and seasonal temperatures. The seasonal temperatures were the 3-month means for December-February (winter), March-May (spring), June-August (summer) and September-November (autumn). The grid-cell trends were then objectively analyzed using the spatial weighting technique of Cressman (1959) with a 400 km radius of influence. Figures 1 and 2 show the analyzed trends of the annual and seasonal temperatures, respectively. The black area in each plot is the central Arctic "data void," which consists of all points at which the data did not permit the evaluation of a trend anywhere in the surrounding 400 km. Even in the offshore areas of the Arctic Ocean, the plotted trends are questionable because they represent extrapolations from the nearest land stations. Despite the data void over the Arctic Ocean, some noteworthy patterns emerge:

- (1) Warming dominates in the winter and spring seasons, as projected by global climate models (IPCC, 1990, Plate 5.4).
- (2) The areally-weighted trend for the summer is nearly zero.

# Arctic Temperature Trends 1961-1990

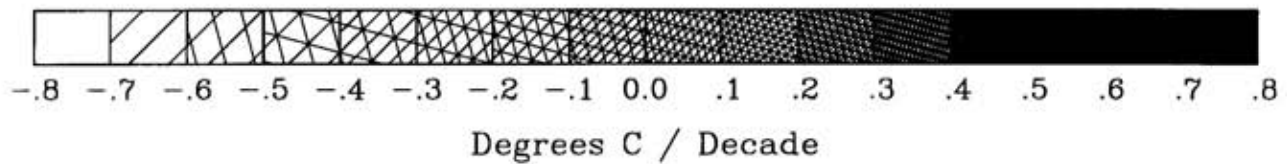
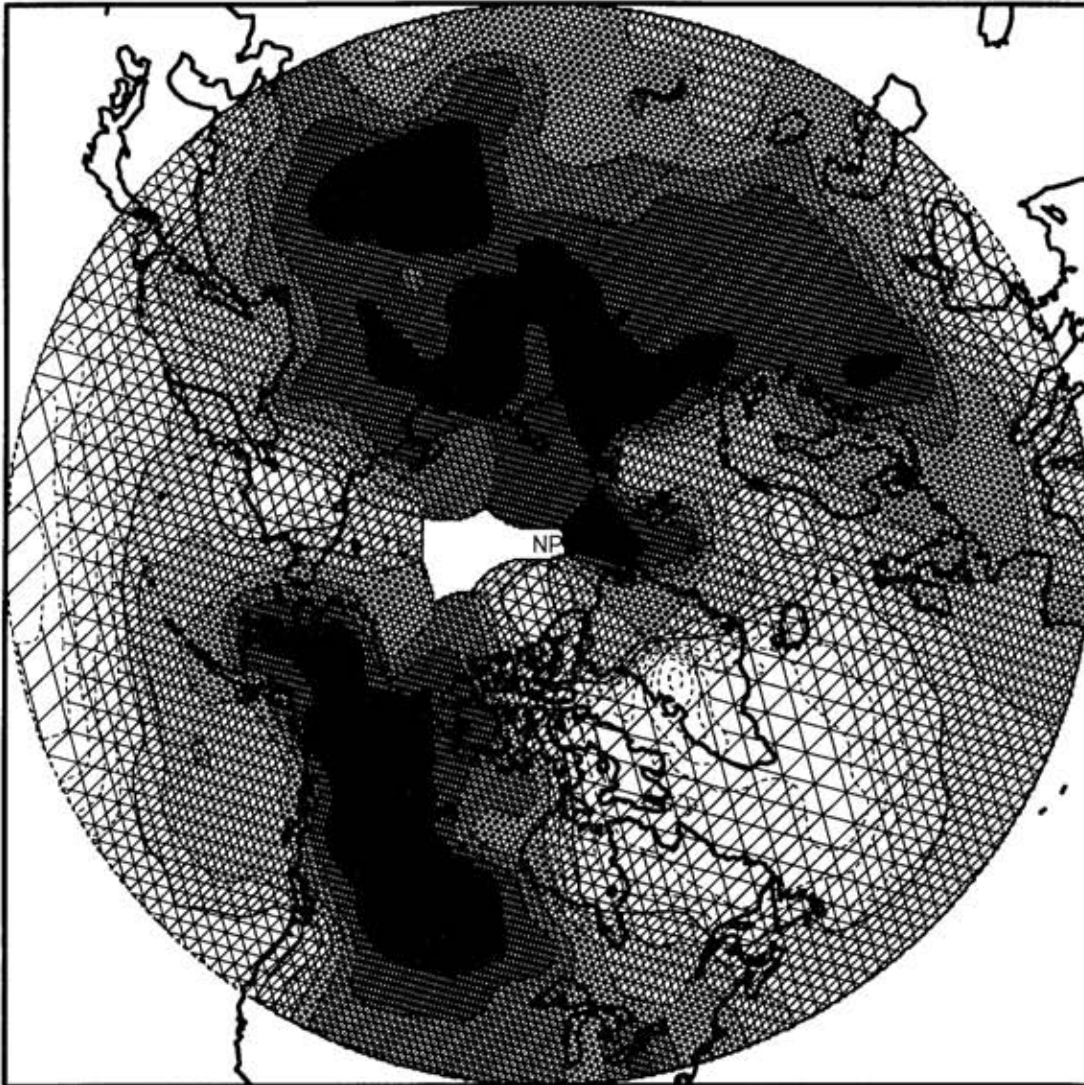
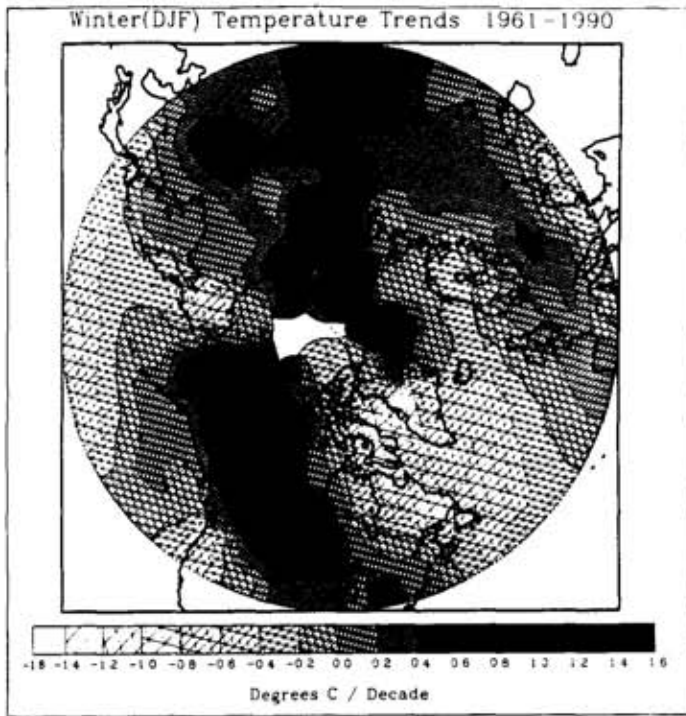
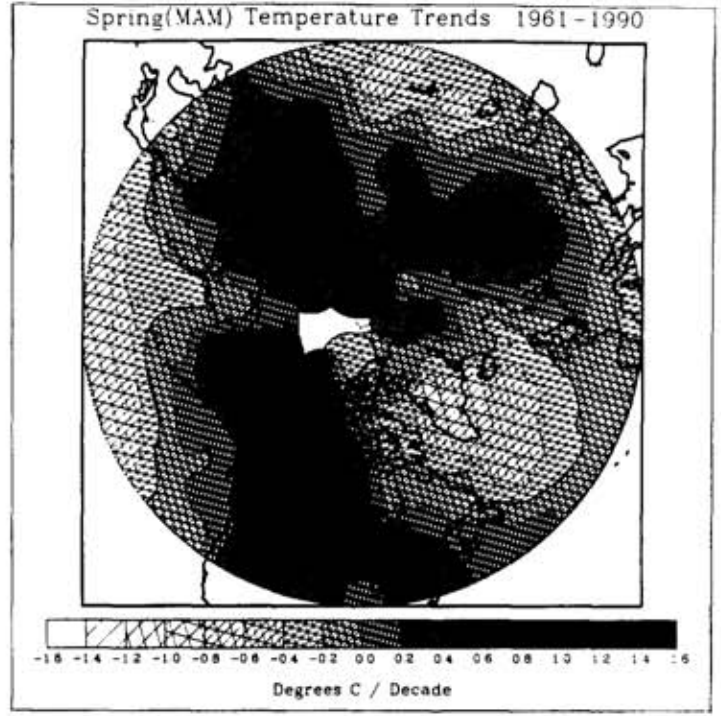


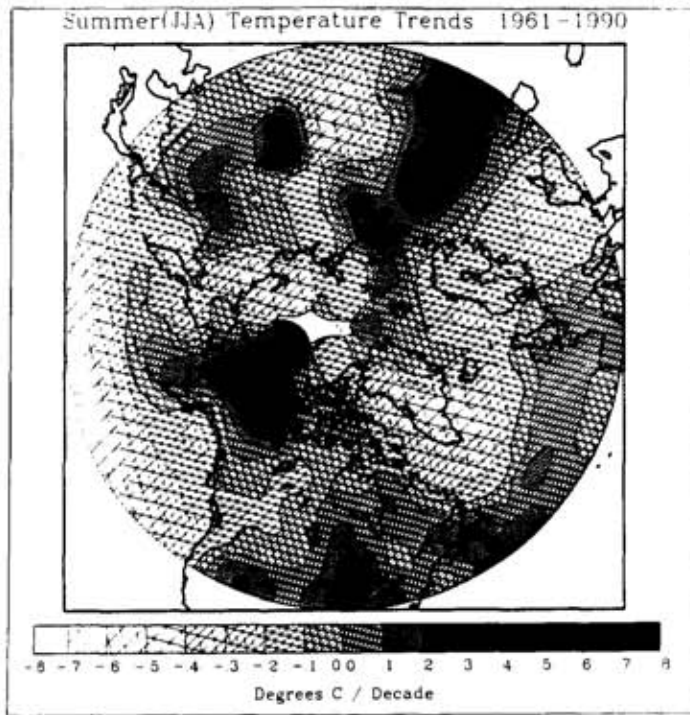
Figure 1. Trends of annual temperature ( $^{\circ}\text{C}$  per decade) computed using linear regression and Cressman analysis procedure with 400 km radius of influence.



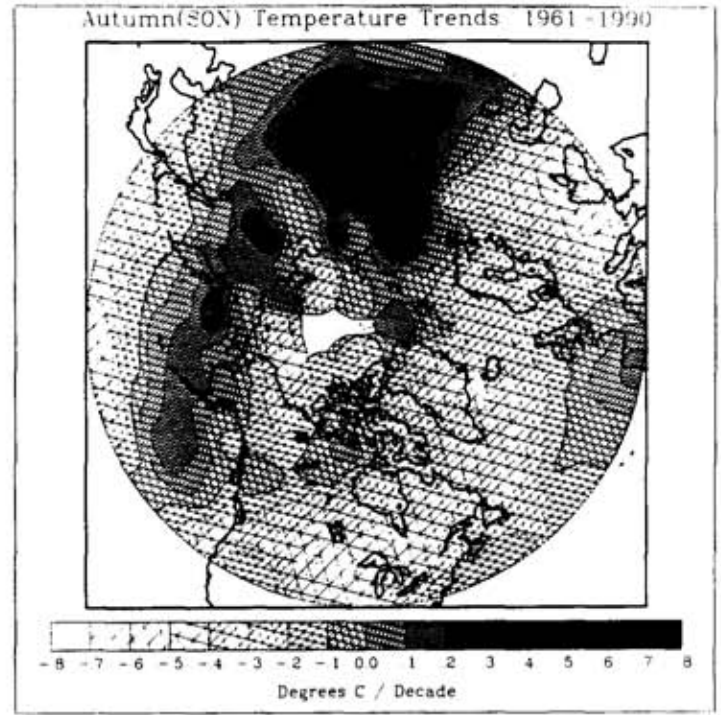
a



b



c



d

Figure 2. As in Figure 1, but for the seasonal temperatures. Seasons are defined here as winter (a. December-February), spring (b. March-May), summer (c. June-August) and autumn (d. September-November). *Note:* Some detail is lost because of the figure size. Larger copies are available on request



- (3) The warming is strongest over the subpolar land areas of Alaska, northwestern Canada and northern Eurasia.
- (4) Little or no warming, and even a slight cooling, is indicated in western sections of the subpolar North Atlantic.

The last two features contribute to a spatial pattern that shows at least some similarity to the greenhouse warming pattern obtained in the more recent experiments with coupled atmosphere-ocean models, which show strong winter warming over the northern land areas and little or no warming in the North Atlantic (Manabe et al., 1992; Washington and Meehl, 1989; UKMO, 1991 [Warrilow, pers. comm.]). The data-derived patterns in Figures 1 and 2 differ from the model-derived patterns in some respects, e.g., the model's projected warmings for North America are generally centered farther east than Alaska and the Yukon. The area of cooling in the central North Pacific is also not found in the model projections. However, this feature is relatively narrow in the north-south direction, extending south of the equatorward boundary of Figures 1 and 2 by only 5-10° of latitude; it is also consistent with the broader field of temperature anomalies of the 1980's (IPCC, 1990, Plate 7.13c).

Time series of the zonally averaged temperature anomalies since 1900 (not shown) indicate that the warming of the early twentieth century was highly amplified in the Arctic, from approximately 60°N to the northern limit of available data at approximately 75°N. However, the recent warming is strongest in the 50-60°N zone and decreases slightly as one goes north of 60°N. This latitudinal dependence of the zonally averaged warming is consistent with Figure 1. Since the more poleward (>70°N) areas contain larger fractions of ocean and sea ice, the latitudinal dependence of the recent warming may be viewed as a manifestation of the tendency for land areas to have warmed the most in a hemisphere in which the land/sea fraction varies considerably with latitude.

Although the patterns in Figures 1 and 2 have some features in common with model-derived greenhouse projections, other factors confound the interpretation of the temperature data. First, the warming in at least some of the high-latitude land areas (e.g., Alaska) is consistent with a shift of the atmospheric circulation toward a pattern of stronger warm advection, especially in the winter season (Walsh and Chapman, 1990). Kalkstein et al. (1991) indeed show that the frequencies of the coldest airmasses have decreased since the 1940's over northwestern North America. However, Kalkstein et al. also present some evidence that the coldest airmasses in this region have warmed by several degrees since the 1940's. One cannot dismiss urbanization as a factor contributing not only to the latter finding but also to the broad wintertime warming of Figure 2. Finally, low-frequency variations associated with the internal dynamics of the ocean-atmosphere system, with tidal cycles, or with solar activity cannot be dismissed as possible explanations of at least some of the apparent changes of northern temperatures over the past few decades.

### *Sea ice*

In order to analyze sea ice variations over the past several decades, we have consolidated the weekly sea ice analyses produced since 1972 by the Navy/NOAA Joint Ice Center (Gross, 1986). We have interpolated the Joint Ice Center (JIC) grids to a 110 km grid and to the ending dates of each calendar month, and we have used regional ice data sources to extend the series of monthly grids back to January, 1953 (Walsh and Johnson, 1979). The data for the pre-1972 period are less homogeneous than are the JIC data because of factors such as (a) differences in the analysis procedures used by the various agencies and (b) the need to temporally interpolate the data for some regions when no data were available. In order to minimize the impact of inhomogeneities in the charting of ice

concentrations, we examine time series of "ice extent" by considering the area covered by sea ice having a concentration of at least 0.1.

Figure 3 is a time series of ice extent for the entire Arctic Ocean and subarctic seas (excluding the Baltic and Okhotsk Seas). Although the variability is dominated by the annual cycle, a slight downward trend is apparent in the summer minima. New minima for the post-1973 period were achieved in individual months of the summers of 1977, 1981 and 1990.

The sea ice data for 1961-1990 have been used in an assessment of the recent trends of total ice extent in the Arctic. The use of this 30-year period eliminates the subset (1953-1960) of ice data most likely to introduce inhomogeneities into the analysis, and it makes the ice data temporally compatible with the temperature data discussed earlier. The 30-year dataset was used to evaluate trends of the annually averaged ice extent and of the seasonal (three-month) averages of ice extent. Figure 4 shows that time series of the ice extent in each of the seasons. The recent decrease of ice extent suggested by Figure 3 is most apparent in the spring and summer time series of Figure 4.

The statistical significance of the trends was evaluated by a Monte Carlo procedure in which the corresponding trends of 1000 randomly reordered versions of each time series were evaluated by linear regression. The slopes of the actual time series of Figures 3 and 4 were placed into the distribution of slopes of the corresponding randomized series, and the probabilities of obtaining by chance the actual slopes (trends) were determined accordingly. The following are the slopes and chance-level probabilities of those slopes:

	<b>Slope</b> <b>(10<sup>6</sup> km<sup>2</sup> per decade)</b>	<b>Change</b> <b>(% per decade)</b>	<b>Probability</b>
Annual	-0.15	-1.2	.046
Winter (JFM)	+0.02	+0.1	>.200
Spring (AMJ)	-0.17	-1.2	.121
Summer (JAS)	-0.35	-3.6	.017
Autumn (OND)	-0.10	-0.8	.164

As indicated by the table, the negative trend of summer is statistically significant at the 2% level. The negative trends of spring and autumn are not, nor is the very slight increase of winter. The negative trend of summer is sufficiently large that it gives the time series of annual (12-month) averages a statistically significant negative trend.

Figure 5 shows the longitudinal distributions of the percentages by which the 15-year mean ice extent changed from 1961-1975 to 1976-1990. The changes for winter (Figure 5a) are quite small except for the eastern North Atlantic, where decreases of 5-10% are indicated. An increase of approximately 8% occurred in the longitudes (300-320°E) of Davis Strait and the Labrador Sea, where the temperature data of Figures 1 and 2 indicated a cooling during 1960-1990. The changes for summer (Figure 5b) are generally negative and are quite large (>10%) in some sectors: the Greenland Sea, the Barents Sea and the Kara Sea. The summer changes in the Pacific portion of the Arctic are between 0 and -5% in most 20° sectors.



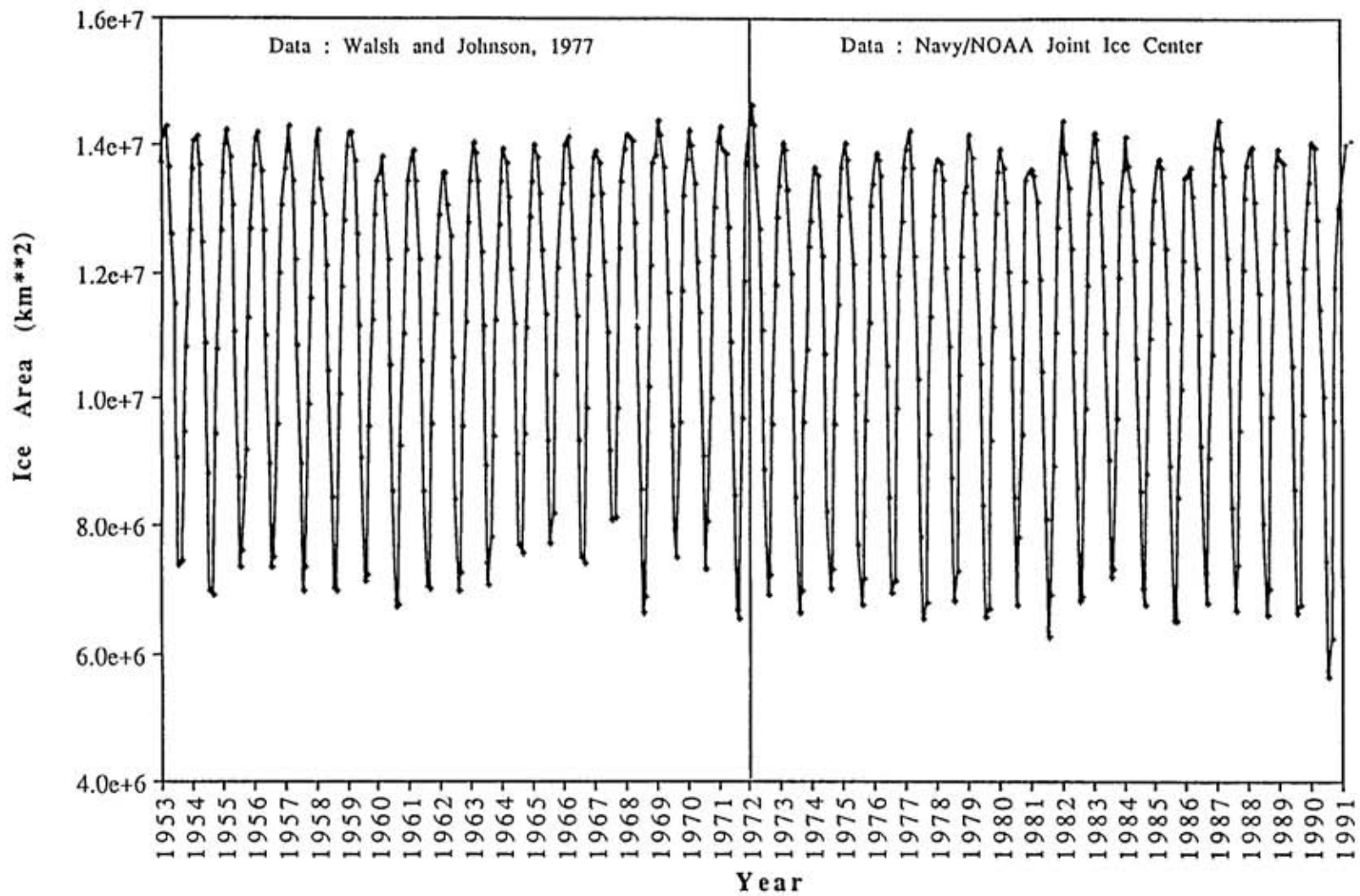


Figure 3. Time series of end-of-month arctic sea ice extent, 1953-1990.

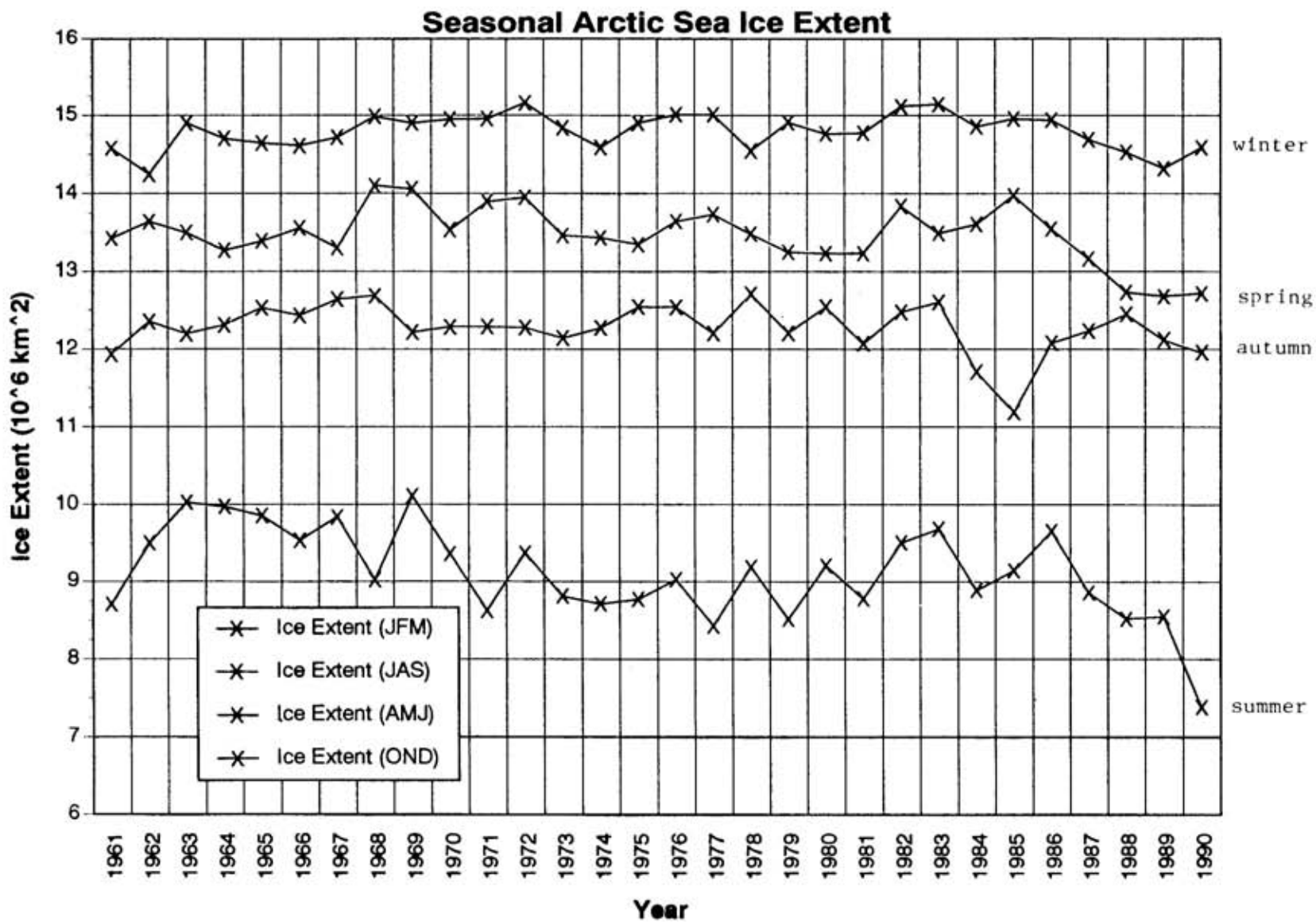
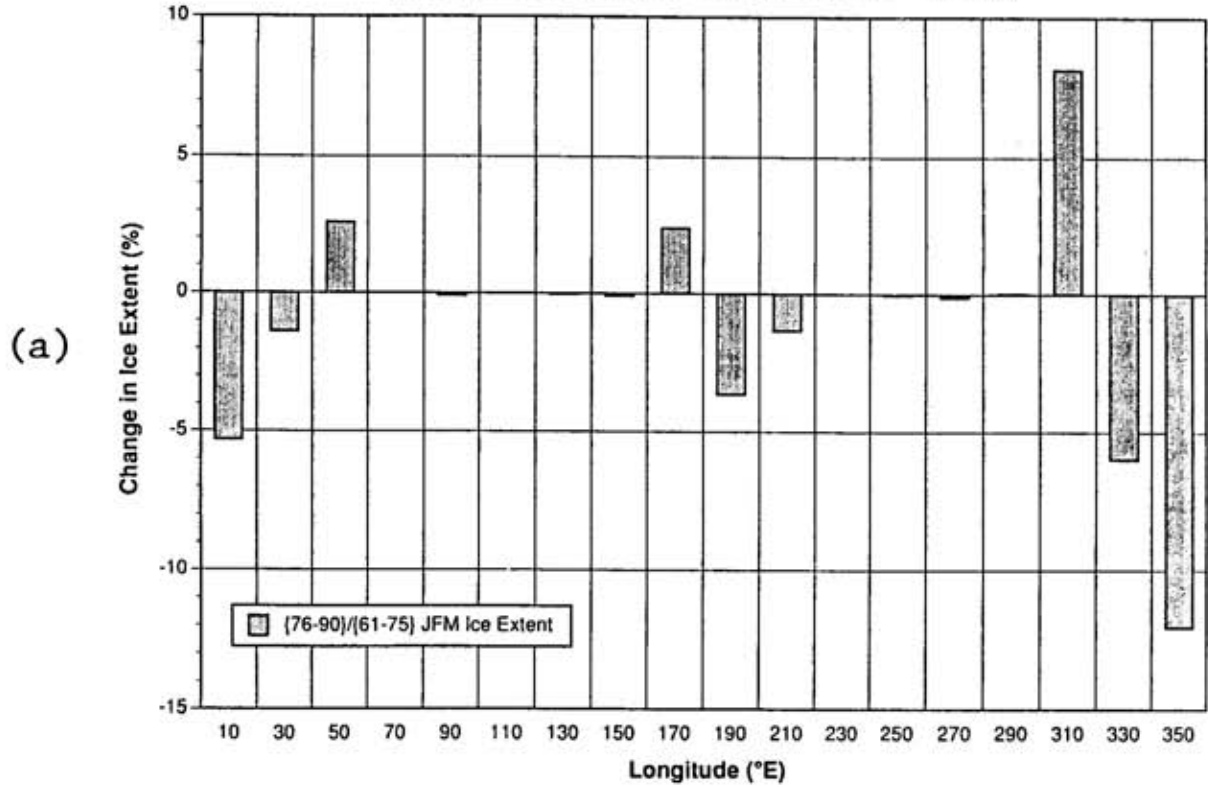


Figure 4. Time series of end-of-month arctic sea ice extent averaged over seasons. From top to bottom, plotted time series are for winter (JFM), spring (AMJ), autumn (OND) and summer (JAS).

Change in Means of Winter(JFM) Sea Ice Extent  
(Change from {1961 - 1975} to {1976 - 1990})



Change in Means of Summer(JAS) Sea Ice Extent  
(Change from {1961 - 1975} to {1976 - 1990})

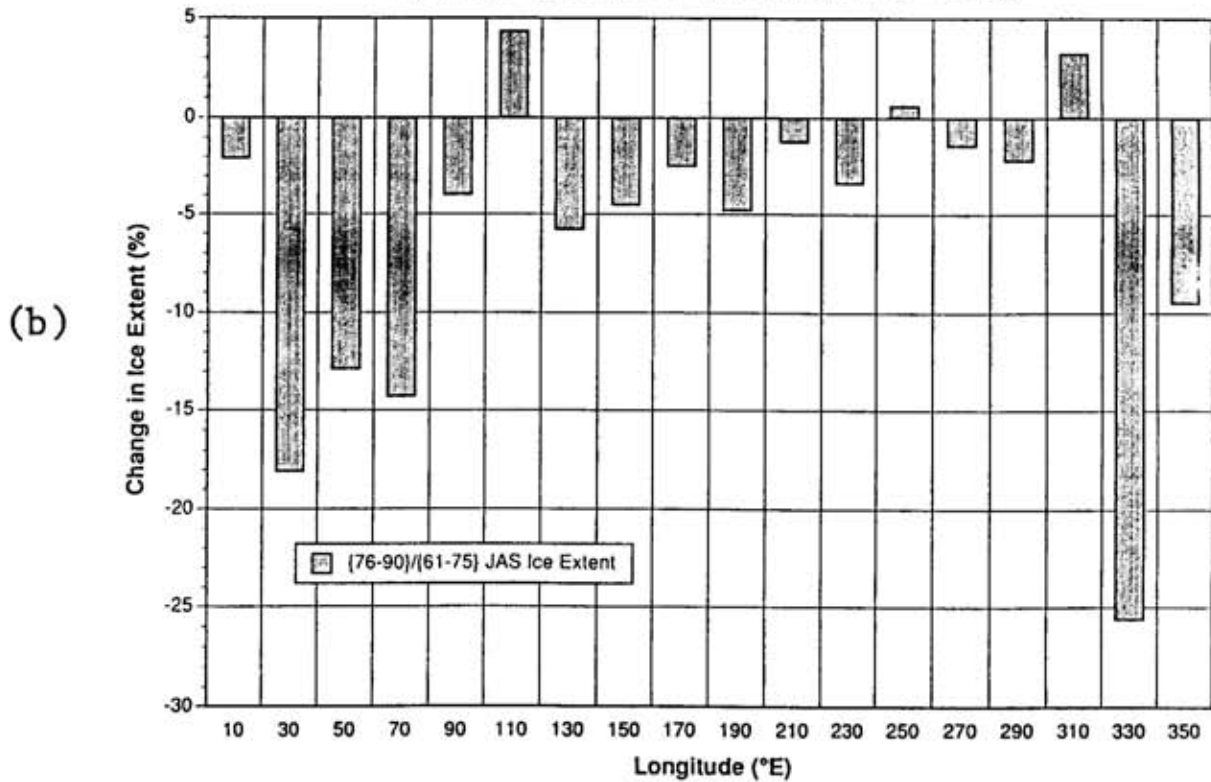


Figure 5. Percentages by which mean sea ice extent changed from 1961-75 to 1976-90. Changes are plotted by 20° longitudinal sector for (a) winter, January-March and (b) summer, July-September.

**Change in Means for Sea Ice and Surface Air Temperatures  
(Change from {1961 - 1975} to {1976 - 1990})**

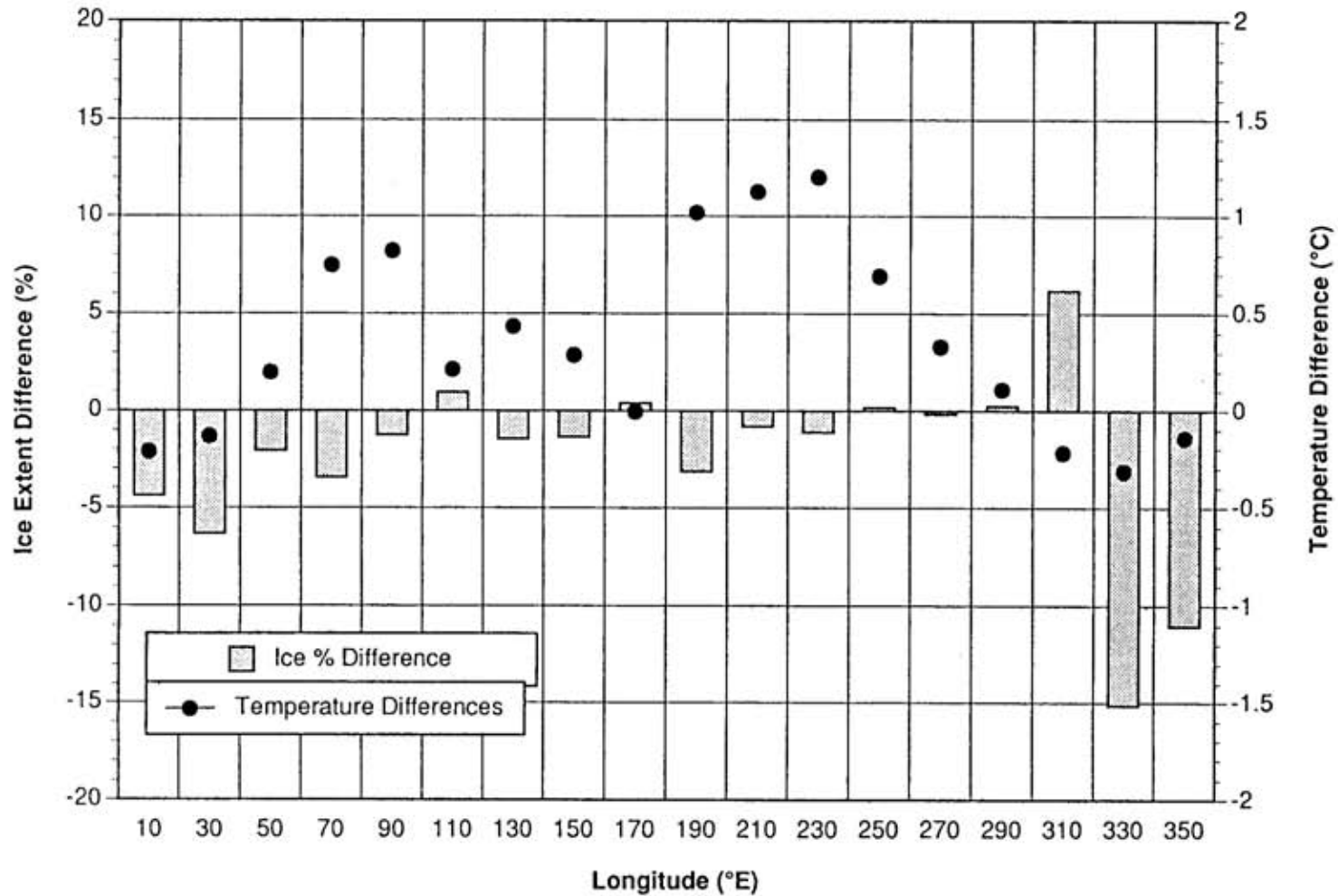


Figure 6. Changes in annually averaged sea ice extent and surface air temperature (55-75°N) from 1961-75 to 1976-90. Changes of ice extent (% , scale on left) are shown by vertical bars; changes of air temperature (°C, scale on right) are shown by solid circles.

Finally, Figure 6 shows the longitudinal distribution of the changes of annually averaged ice extent and air temperature (55-75°N) from 1961-75 to 1976-90. In general, the changes of ice extent and air temperature are of opposite sign. An exception is the North Atlantic subarctic (40°W-40°E), where the data indicate a decrease of ice despite a slight cooling in the southern portion of that region. The largest positive changes of temperature occur in those longitudinal sectors (60-160°E, 80-180°W) in which the winter advance of the ice is limited by the continents. This constraint effectively reduces the percentage by which the annually-averaged ice extent can change because the only period of major variability (or change) is the summer portion of the year.

## Conclusion

The data for the past several decades indicate a general warming of much of the Arctic and subarctic. The warming of the annual mean temperature is strongest over the land areas and is determined mainly by the patterns of temperature change for the winter and spring. The warming is generally compatible with recent changes of sea ice, for which the summer extent has decreased substantially. As discussed in other papers at this Workshop, there are indications that continental snow cover has also decreased during the past few decades (e.g., Robinson, this volume). Close monitoring of these variables during the next few years appears to be highly advisable in order to determine whether the recent variations are short-term (several-year) excursions from relatively stable means or whether they are indicative of more systematic changes.

Attention should also be given to data uncertainties, particularly in the high-latitude temperatures, which suffer from undersampling in the subpolar seas and from possible urban-induced contamination in the northern land areas.

Finally, the results summarized here are not inconsistent with greenhouse scenarios projected by some climate models. Before a convincing greenhouse linkage can be established, however, one must address (1) the model-to-model differences between various coupled atmosphere-ocean model experiments, and (2) the possibility that alternative sources of low-frequency variability may be responsible for the recent variations described here.

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# Inferences of the North American Snowfall and Snow Cover With Recent Global Temperature Changes

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## Introduction

Climate models and paleoclimatic reconstructions suggest that a global warming, projected due to anthropogenic increase of the greenhouse gases ( $\text{CO}_2$ ,  $\text{CH}_4$ ,  $\text{NO}_x$ , etc.) in the atmosphere, should manifest itself most strongly in the high latitudes (Budyko and Izrael 1987, IPCC 1990). According to the scenarios of this warming, an increase of mean temperature and total precipitation in high latitudes will occur. This could affect the Northern Hemispheric snow cover areas in different ways. The warming and increased precipitation as rainfall will force a retreat of the snow cover. However, the increasing snowfall on the contrary will provide additional sources for snow cover extension. So, a priori, the sign of the future changes of snow cover in high latitudes is uncertain. This is a primary reason to analyze the empirical data on contemporary changes in a snow cover expansion in an attempt to reveal its possible relationships with the current changes in global temperature and high latitude precipitation.

This study of regional changes in snow cover is restricted to the analysis of data for North America. The Eurasian snow cover data will be considered only on a continental scale. Satellite snow cover data (Matson 1986) that were updated and corrected in the U.S. Climate Analysis Center (Ropelewski 1991) were used for the period Sept.1973 - Aug.1991. The corrections used are consistent with ones proposed by Robinson et al. (1991). Annual snowfall and total precipitation data for approximately 1300 U.S. and Canadian stations were used for the period 1950-1990 (for Canada up to June 1990).

## Continental scale relationships between snow cover and Northern Hemisphere temperature.

The presence of relationships between hemispheric snow cover expansion and surface air temperature was shown by Robinson et al. (1991). In this paper these relationships are studied on the continental and (for North America) regional scales.

Annual (Sept.-Aug.) area-mean North American and Eurasian snow cover area ( $S_A$  and  $S_E$ ) are presented in Fig. 1. Mean annual (Jan.-Dec.) temperature anomalies of the Northern Hemisphere by Vinnikov et al. (1990), ( $T$ ) are plotted on the same graph. The time series are negatively correlated. The correlation is relatively weak ( $R=-0.5$  for both

# Northern Hemisphere temperature anomalies and mean annual snow cover area over:

(a) Northern America, (b) Eurasia

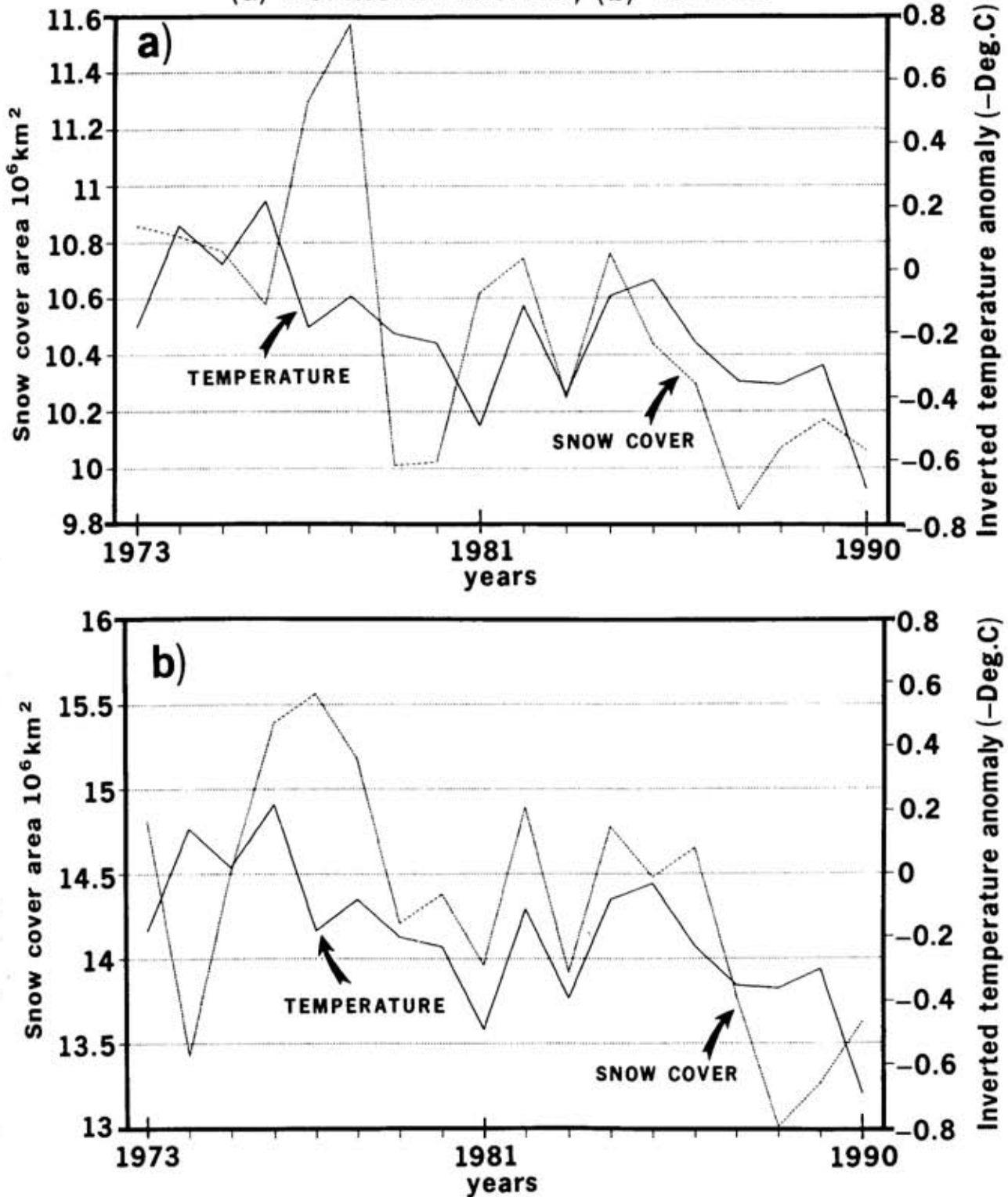


Figure 1. Variations of annual (Sept.-following Aug.) (a) North American and (b) Eurasian snow cover areas and the Northern Hemispheric surface air temperature annual (Jan.-Dec.) anomalies by Vinnikov et al. (1990), 1973-1990.

$S_A$  and  $T$ , and for  $S_E$  and  $T$ ), but statistically significant at the 95% confidence level. Regression analysis indicates that the annual change of snow cover with respect to temperature is  $-1.0 \cdot 10^6 \text{km}^2/\text{°C}$  for North America and  $-1.6 \cdot 10^6 \text{km}^2/\text{°C}$  for Eurasia.

Regression provides biased (low) estimates of the functional relationship when both variables contain errors (Kendall and Stuart 1967). To avoid these biases we used another approach to estimate  $dS/dT$ . If the 18-year period record (1973-1990) is divided into the nine warmest and nine coolest years of the Northern Hemisphere, it can be seen that these years usually correspond with low-snow and high-snow years, respectively (Table 1). The grouping method (Wald 1940) based on the data from Table 1 gives estimates of the  $dS/dT$ :  $dS_A/dT = -2.2 \cdot 10^6 \text{km}^2/\text{°C}$  and  $dS_E/dT = -2.8 \cdot 10^6 \text{km}^2/\text{°C}$ . These are about twice the magnitude of the regression values and have 95% confidence intervals of  $(-3.9, 1.3)$  and  $(-5.8, -1.0)$  respectively. This result shows that the regression estimates of  $dS/dT$  could be considered as conservative.

**Table 1.** Mean values of the Northern Hemisphere annual (January-December) temperature anomalies (Vinnikov et al. 1990) and annual (September-following August) continental scale snow cover (Ropelewski 1991) with partition of the data into two groups of the nine warmest and nine coolest years for the period 1973-1990.

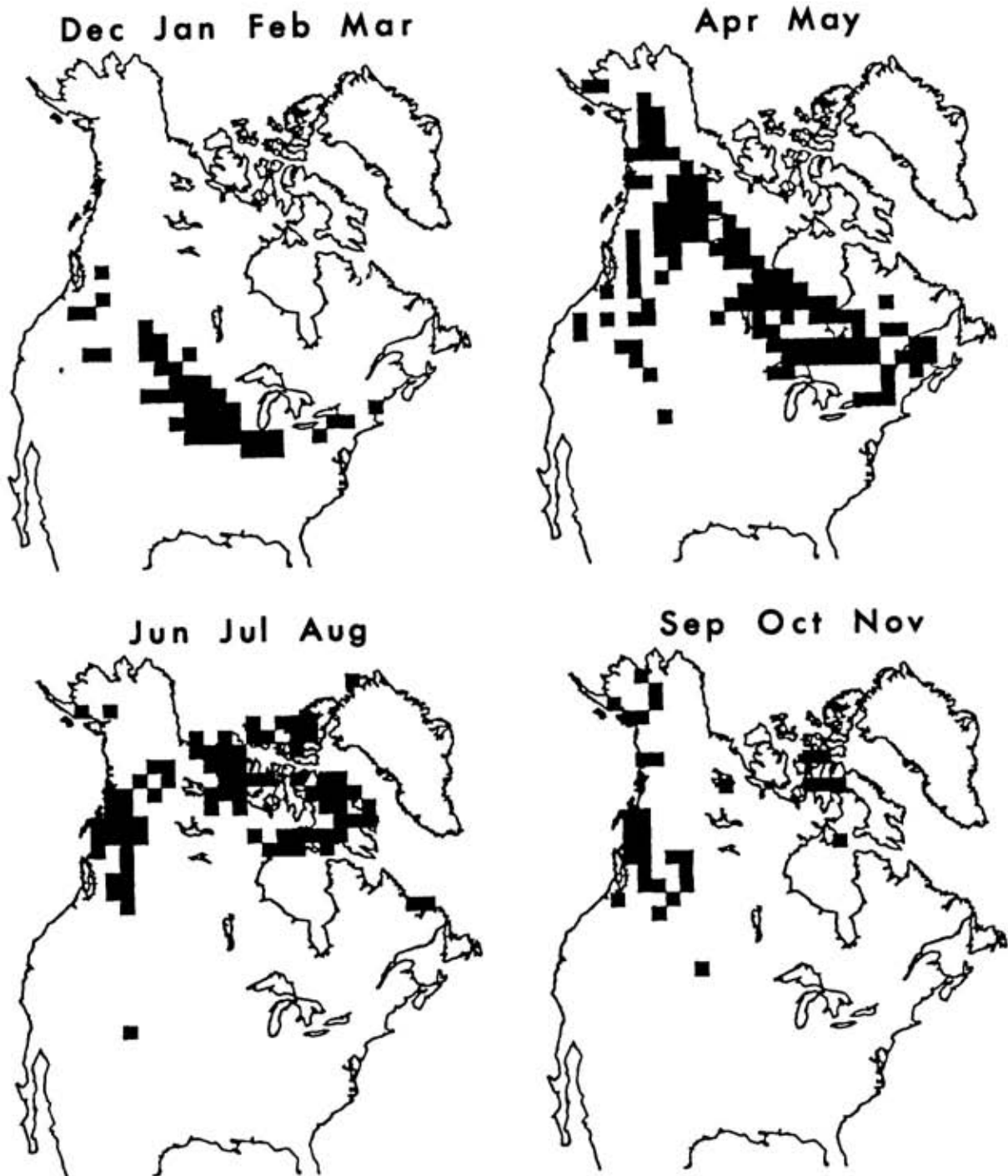
Period, group of years	Temperature anomaly $^{\circ}\text{C}$ (from the period 1951-1975)	Continental snow cover ( $10^6 \text{km}^2$ ) over	
		Northern America	Eurasia
1973-1978, 1982, 1984-1985	0.03	10.9	14.8
1979-1981, 1983, 1986-1990	0.36	10.15	13.9

When the estimates of  $dS/dT$  are used in conjunction with scenarios of  $0.5^{\circ}\text{C}$  warming in the global surface air temperature to the end of next decade (Budyko and Izrael 1987), a 5 to 10% decrease of the annual North American and Eurasian snow cover results.

In Karl et al. (1992) we defined the seasonal responsive zones for the North American continent, where the main changes in the snow cover occurred between the warm low-snow and cold high-snow years. To obtain these zones we used the same partition (Table 1) of the 18-year period (1973-1990), into the nine warmest and nine coolest years. Fig.2 depicts the regions, where the average probability of presence of the snow cover decreased to a minimum of 15%, when the hemispheric temperature increased by 0.33 degree Celsius. In this case, the regions shown in Fig.2 and their northern vicinities would be the first areas, where the projected global warming would affect the seasonal expansion of the snow cover.

#### **North American zonal changes in snowfall**

Fig.3 depicts three area-mean annual snowfall time series (zones  $35 - 45^{\circ}\text{N}$ ,  $45-55^{\circ}\text{N}$ , and  $55-70^{\circ}\text{N}$  (without Alaska)). The time series were obtained by the Thiessen method of spatial averaging (1911) of the data from about 1300 stations (404, 955, and 105 for each zone respectively). Data from the stations located in the one degree latitudinal belt along the border between zones were used twice in spatial averaging (once for both zones) but with



**Figure 2.** Map of the most responsive regions of snow cover differences between cold, high-snow years and warm, low-snow years. Monthly snow cover data were gridded, composited by season, and averaged separately for the two groups of years. The gridbox is displayed when the difference exceeds 15%.

low weights. To avoid the influence of missing data, the data were transformed into anomalies from the reference period 1971-1990 prior to averaging.

Fig 3 shows that annual snowfall over northern Canada (55 to 70 degrees north) has increased during the last four decades. The North American meteorological network north of 55 N was inadequate for detailed analysis until after World War II. Therefore, this increase in snowfall (and in rainfall) characterizes the whole instrumental period of meteorological observations in the region. It is unwise to extrapolate trends without a physical explanation of their origin. Therefore, we cannot predict future precipitation changes in this zone. However, the reasons for this constant increase (corresponding with the century-scale increase of annual precipitation over the northern part of Eurasia (Groisman 1991) have to be carefully investigated. This is one of the greatest large-scale trends in precipitation observed during the instrumental period.

The area-averaged North American annual snowfall over the zone 45-55°N is negatively correlated with the annual surface air temperature over the Northern Hemisphere ( $R=-0.7$ ). These two time series are depicted in Fig.4 with an inverted scale for the hemispheric temperature anomalies. The inference from this relationship is that during a projected global warming, the snowfall over this zone will likely decrease.

No trends in the North American area-averaged annual snowfall over the zone 35-45°N were found. The snowfall in the 1950s and 1980s was below normal (i.e. below the four-decade average), and in the 1960s and 1970s was above it.

Table 2 lists the statistical characteristics of the snowfall time series depicted in Fig.3. The mean values and standard deviations are presented separately for two periods: the 1980s and the previous three decades (1950-1979). The differences in snowfall totals between these two periods are statistically significant at the 95% level only for the northern Canada annual values. In the 1980s, northern Canada obtained an excessive amount of snowfall (+10%) while the zone from 45-55°N experienced a 6% reduction in annual snowfall. The annual snowfall in the zone 35-45°N is not correlated with the hemispheric temperature, and during the 1980s it was not significantly different from the previous decades.

This analysis of the contemporary trends in snowfall suggests that there would be no additional snowfall to the south of 55°N to support an expansion of snow cover during a global warming. The snowfall totals have decreased to the south of 55°N during the last (warm) decade. Therefore, the retreat of the snow cover in these regions would likely accompany a projected global warming.

**Table 2.** Statistical characteristics of the North American annual zonally averaged snowfall time series. Long-term mean values and standard deviations are presented for the 1980s and previous three-decadal period (1950-1979) separately. The Canadian data were available only up to June 1990.

Period (years)	North American annual snowfall area-averaged over three latitudinal zones. Mean values and standard deviations respectively (cm).		
	Canada, 55-70°N	45 - 55°N	35 - 45°N
1950-1979	166 14	222 22	97 12
1980-1990(1989)	181 9	208 24	93 14



# Annual North American zonal snowfall stick measurements (cm)

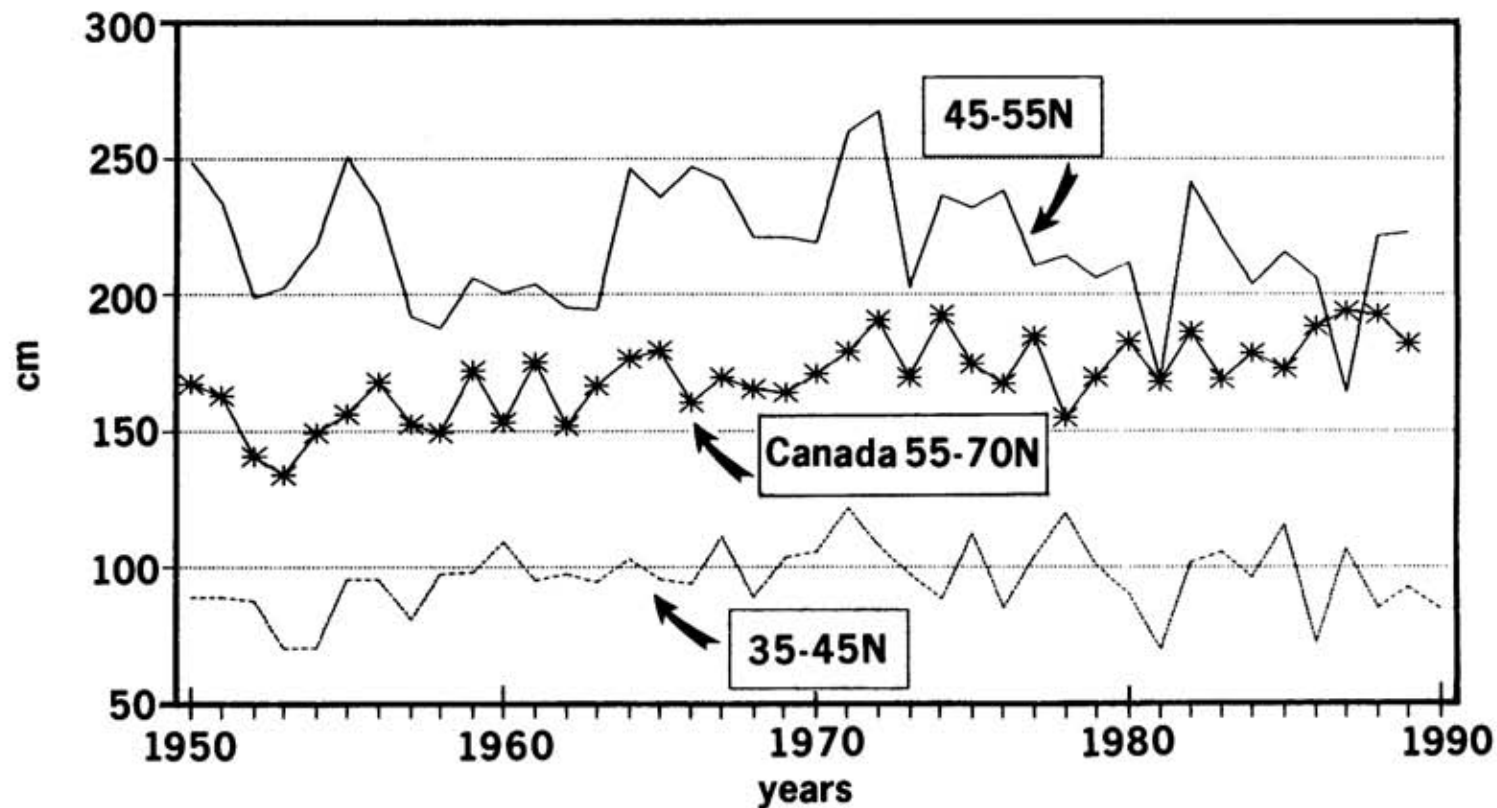


Figure 3. Annual (Jan.-Dec.) North American snowfall, area averaged over the three latitudinal zones: 35-45°N, 45-55°N, and 55-70°N (only Canada), 1950-1990.



# North American snowfall in 45-55 N (S) and hemispheric temperature anomalies (T)

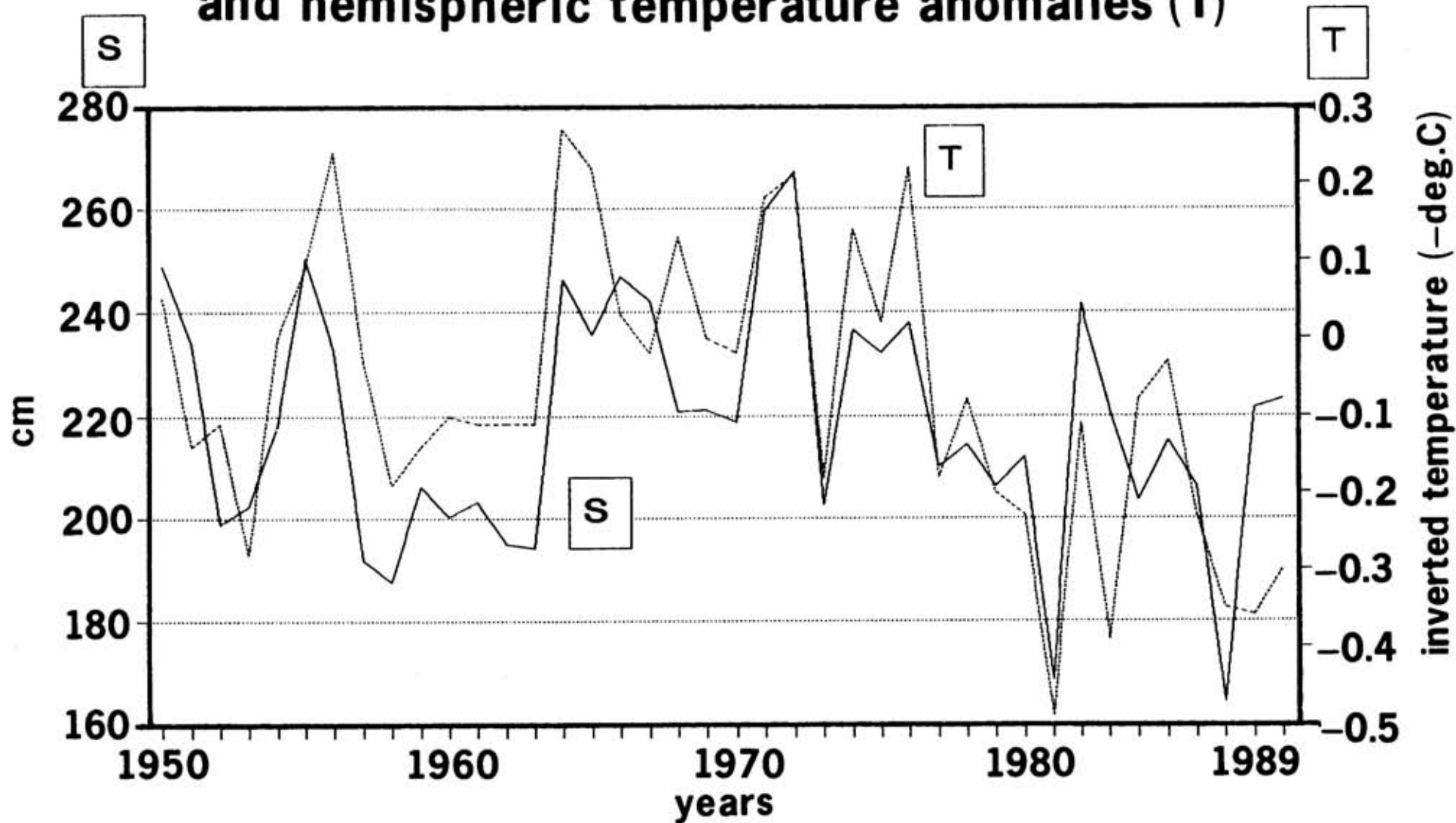


Figure 4. Annual (Jan.-Dec.) North American snowfall, area averaged over zone 45-55°N and mean annual hemispheric surface air temperature by Vinnikov et al. (1990)

In contrast, the annual snowfall increased during the last four decades over northern Canada, with a mean rate of  $4.7\% \pm 0.8\%$  per decade. In order to estimate the future expansion of the snow cover area in the North American region to the north of  $55^{\circ}\text{N}$ , a more complicated analysis has to be done. This future analysis has to include the scenarios (empirical estimates or modelling results) of the changes of the regional spring and summer surface air temperature. The rates of changes of temperature and snowfall will determine the future of the high latitudinal snow cover.

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# Detection of Fluctuations in Global Snow Cover Using Passive Microwave Remote Sensing

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Snow cover is an important variable for climate and hydrologic models due to its effects on energy and moisture budgets. A primary factor controlling the amount of solar radiation absorbed at the earth surface is the extent of snow cover due to its high albedo. Any decrease in snow cover resulting from a warming trend results in increased absorption of solar radiation and additional heat to melt increased amounts of snow. This results in the classic positive temperature-albedo feedback mechanism which is included in nearly all climate models. In addition to the albedo effect, snow cover represents a significant heat sink during the warming period of the seasonal cycle due to a relatively high latent heat of fusion. As a result, the seasonal snow cover provides a major source of thermal inertia within the total climate system as it takes in and releases large quantities of energy with little or no fluctuation in temperature. Therefore, snow cover is an important variable for global climate monitoring and change detection.

During the past two decades much important information on hemispheric-scale snow extent has been provided by satellite remote sensing in the visible wavelengths (Robinson et al., 1991; Dewey, 1987). One major problem with the application of visible wavelength data is the need for clear sky conditions. Persistent cloud cover in winter may limit data to only a few days per month. In addition, visible-band data do not provide the opportunity to extract snow water equivalent or depth information.

Passive microwave remote sensing offers several advantages over visible-band data. It is currently the best method to monitor temporal and spatial variations in snow cover on hemispheric to global scales. Passive microwave remote sensing allows data collection in nearly all weather conditions and during darkness, and also provides the potential to compute snow water equivalent and to detect melt (Kunzi et al. 1982; Foster et al. 1984; Rott, 1987). These additional variables comprise important input to energy budget, hydrologic, and global circulation models.

Microwave energy emitted from the surface of the earth is measured by the Defense Meteorological Satellite Program (DMSP) Special Sensor Microwave Imager (SSM/I) which provides data in four separate channel frequencies (Weaver et al., 1987). SSM/I coverage is global and frequent, at least daily for locations above 43 degrees latitude, and once every two to three days for those areas where snow might be expected at latitudes below 43 degrees. When snow covers the ground, some of the microwave energy emitted by the underlying soil is scattered by the snow grains. When moving from snow-free to snow-covered land surfaces, a sharp decrease in emissivity provides a nearly unambiguous indicator of the presence of snow. The amount of scattering is related to both the amount of snow (number of grains) and specific wavelength. Based on this relationship, algorithms have been developed which indicate the presence of snow and compute snow water equivalent, or snow depth given an assumed density (Chang et al. 1981; Chang et al. 1987a and 1987b; Kunzi et al. 1982; Hallikainen and Jolma, 1986; MacFarland et al. 1987; Goodison, 1989; Aschbacher, 1989; Josberger et al. 1990; Rott et al. 1991; Grody, 1991).

There are also disadvantages to the passive microwave data. The resolution is coarse when compared to that typically available from visible-band sensors (Rott, 1987). Resolution depends on the channel frequency and is approximately 30 km for those channels used in the snow cover algorithms. In addition, the application of passive microwave is limited to dry snow. Once liquid water is present on the ice grains, the snow surface becomes a strong emitter and is indistinguishable from bare ground. This dramatic change in signal can be used to monitor the onset of melt, but as long as the snow is wet, its location and water equivalent cannot be determined using SSM/I data alone. If the snow refreezes, it again become "visible" to the microwave sensor. Another problem results from the fact there may be objects which are primarily emitters which extend above the snow cover, such as dense coniferous forests or bare rock outcrops. The increased emission from such surface features will tend to reduce the total amount of scattering integrated across the pixel and thus may indicate a lesser amount of snow cover than is actually present.

Finally, there are complications which result from the snow structure itself. The amount of microwave scattering is not only dependent on the number of grains, but also on the size of the grains, the larger the grains the greater the scattering. In theory, brightness temperature is very sensitive to grain size (Chang et al., 1981). In order to make any practical use of a given algorithm, the snow grain diameter to within about 0.2 mm would be required. Except for local scale studies, it is simply not be feasible to obtain such detailed snow structure information. However, when existing algorithms are tested the results do not show such a high degree of sensitivity to grain size. Algorithms which assume a single grain size provide reasonable results on the regional scale (Chang et al. 1987b; Kunzi et al. 1982; Aschbacher, 1989; Goodison, 1989; Hallikainen and Jolma, 1986). Apparently the shapes of the natural snow grains are not as sensitive to scattering as the spheres used in the radiation transfer theory. In addition, for regional scale studies, it appears that only the mean grain size may be required and not the detailed layer by layer structure of the snow cover. Although not as sensitive as theory might indicate, testing has shown that algorithms which assume a single grain size (for example 0.6 mm diameter, Chang et al., 1987b) do eventually produce unreasonable values when the mean grain size of the snow cover exceeds about 2.0 mm in diameter (Armstrong, 1990a; Hall et al., 1986).

As long as the snow cover remains sub-freezing there is only one process, temperature-gradient or kinetic metamorphism, which will cause the mean grain size of a layer to exceed a diameter of 2.0 mm (Armstrong, 1985; Colbeck, 1991). Using a relationship between average snow depth for a given region and current depth and temperature data, we are investigating a method which will identify the conditions which result in a snow cover with a mean grain size larger than 2.0 mm. Given this information, a technique to adjust the microwave algorithms for grain size is being developed (Armstrong et al. 1992).

NSIDC is currently involved in two specific projects which apply passive microwave remote sensing data to the study of snow cover. The first project is funded by NASA's Interdisciplinary Research Program to develop a capability for the production of daily snow parameter products from the DMSP SSM/I. A data system is being developed which will produce, archive, and distribute validated snow cover products for community use. Initial emphasis is on Northern Hemisphere snow extent. We are also exploring the potential of the SSM/I for mapping other snow cover properties such as snow water equivalent, snow depth, and dry/wet snow boundary (Armstrong, 1990a).

Within this project NSIDC coordinates the activities of the SSM/I Products Working Team (SPWT) which is a multi-agency and multi-disciplinary working group focussing on



the problems associated with extracting land surface (primarily vegetation, soil, and snow cover) information from SSM/I. Currently, emphasis is on developing optimal binning and gridding routines as well as the selection of one or more snow cover algorithms for use in the distribution of standardized data sets by NSIDC. Snow cover algorithm comparison is being undertaken in a cooperative effort with scientists at the University of Innsbruck, Austria (Rott, et al., 1991), the Canadian Climate Center (Goodison, 1989), and NASA Goddard Space Flight Center (Chang et al., 1987a and 1987b; Chang et al., 1991). Regional test areas selected are the western United States, central Canada, and Europe. In the U.S. the accuracy of the algorithms is being tested by comparison with several validation data sets including snow depth measurements from the National Weather Service and the Soil Conservation Service, as well as output from the prototype Air Force Global Weather Central (AFGWC) snow depth model described below. Later stages of this project will explore the combined research potential of SSM/I-derived snow cover and sea ice products for climate dynamics and global/regional hydrology.

The second NSIDC project utilizing passive microwave data involves the development of a prototype snow depth model to replace the current model used by the U.S. Air Force Global Weather Central (AFGWC) (Armstrong and Hardman, 1991a, 1991b). The initial goal of the project was to analyze weaknesses in the current version of the model, identify enhancements, and design and demonstrate an improved software system. The prototype model which was completed in 1991 provides a state-of-the-art integration of all snow cover data available at AFGWC in order to provide a global snow cover product at a 40 km grid resolution. The basic data generated for each grid point include calculated average and maximum snow depth, age in days of the total snow cover, and days elapsed since the last snowfall, along with appropriate data source flags and summary diagnostics. The model represents the integration of surface and satellite observations. Surface measurements used are from the World Meteorological Organization (WMO) synoptic data collection network. In locations where surface measurements are inadequate, the model applies algorithms which rely on SSM/I data to provide snow cover extent and information on snow depth. Because no single snow depth algorithm is suitable for global scale application, NSIDC is evaluating individual algorithms, as described above, to determine which are most accurate for a given surface-type condition.

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# Lake Ice Observations in Finland in the 19th and 20th Century: Any Message for the 21st?

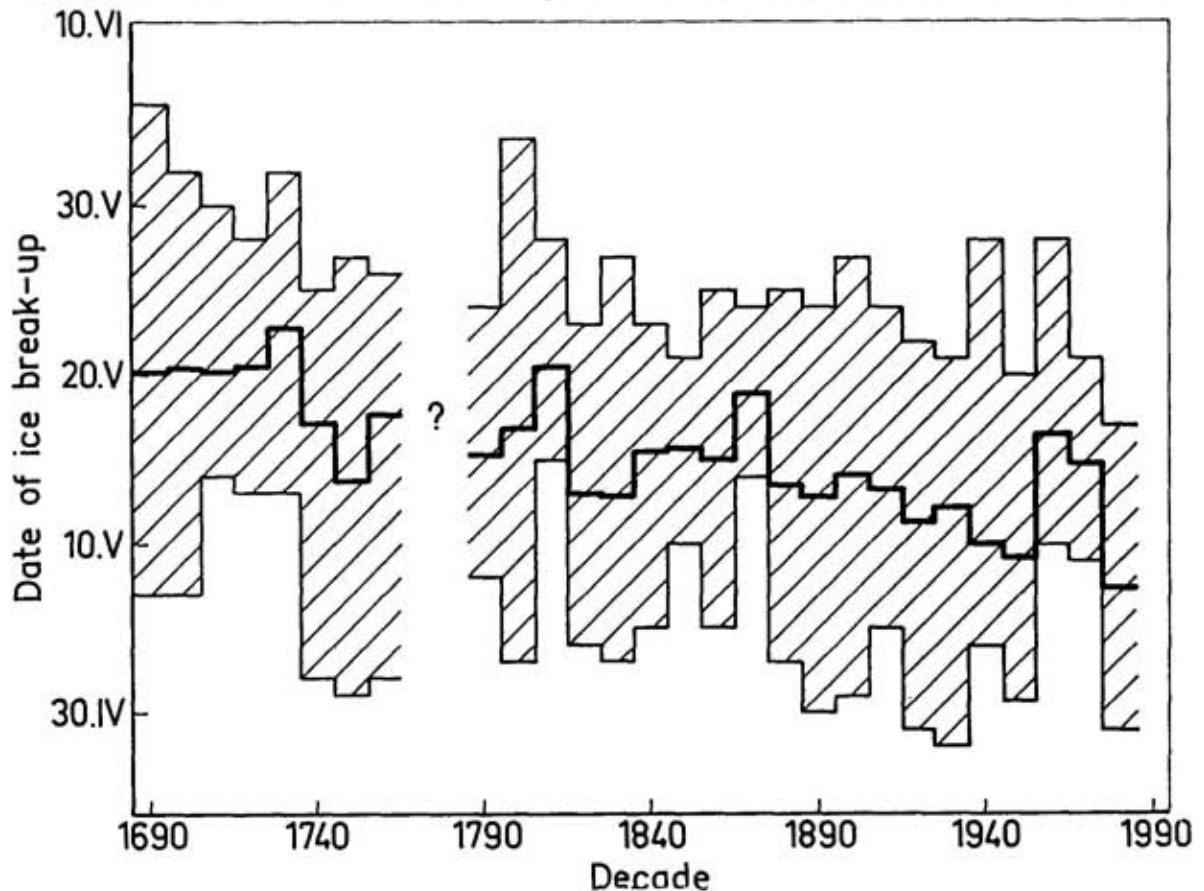
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## Introduction

The Julian calendar was abandoned in Finland in the year 1753. March the 1st was followed by March the 12th. In the River Tornionjoki in Lapland, the break-up in 1752 had occurred on April the 25th. It was a rather early break-up, still unbeaten if it had been a Gregorian date. However, a correct date should naturally be given as May the 6th.

The switch of the calendar is an easily avoidable trap for an interpreter of old hydrological data. There are many other traps, which are far more difficult to avoid. A careful consideration and sound mistrust is always needed.

Figure 1 shows the decennial means and extremes of the Tornionjoki break-up series, which was started in 1693. It is the longest known hydrological data series in Finland.



**Figure 1.** The decennial means and extremes of the break-up date of River Tornionjoki at the town of Tornio.

Several other observation series related to ice phenomena were started during the Julian calendar. However, most of them are too fragmentary to have any special value.

The Little Ice Age can clearly be seen in Tornionjoki data; the break-ups were considerably late from the 1690's to the 1730's. Thus the River Tornionjoki enjoyed the cool conditions two decades past the 'official' end of the Maunder minimum.

Another interesting feature is the earliness of the break-ups in the 1980's. Even the latest break-up in that decade occurred only two days later than the average of the whole series.

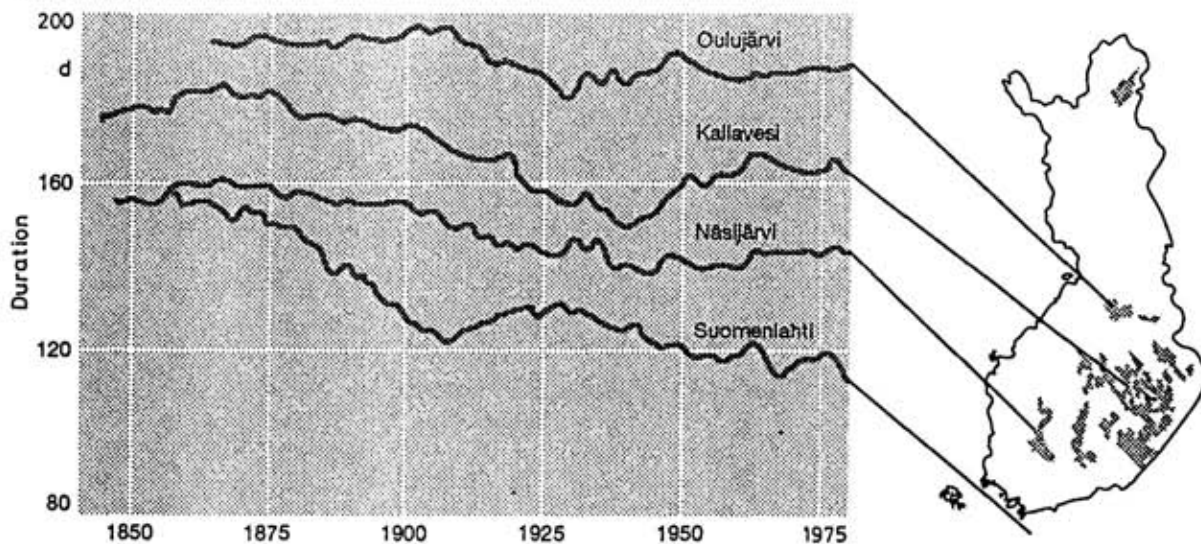
### Lake Ice Observations

The use of freezing, break-up and duration series of lake ice as climatic indices has been more extensive than the use of river ice observations. The lake ice series can also be considered as more reliable indices for two reasons:

- the interpretation of the freezing and break-up dates is easier in lakes than in rivers
- the effect of human activities on lake ice can often be neglected.

Examples on the use of lake ice data in climatic studies are the papers by Tanaka & Yoshino (1982), Palecki & al. (1985) and Tramoni & al. (1985). In Finland e.g. Simojoki (1940), Laasanen (1984) and Kuusisto (1987) have studied this topic. The methods applied in all these papers have been mainly statistical. Usually some variables describing the ice conditions are related to rather arbitrarily chosen variables describing the air temperature. Relatively little attention has been paid to the physical significance of the meteorological variables included in the model.

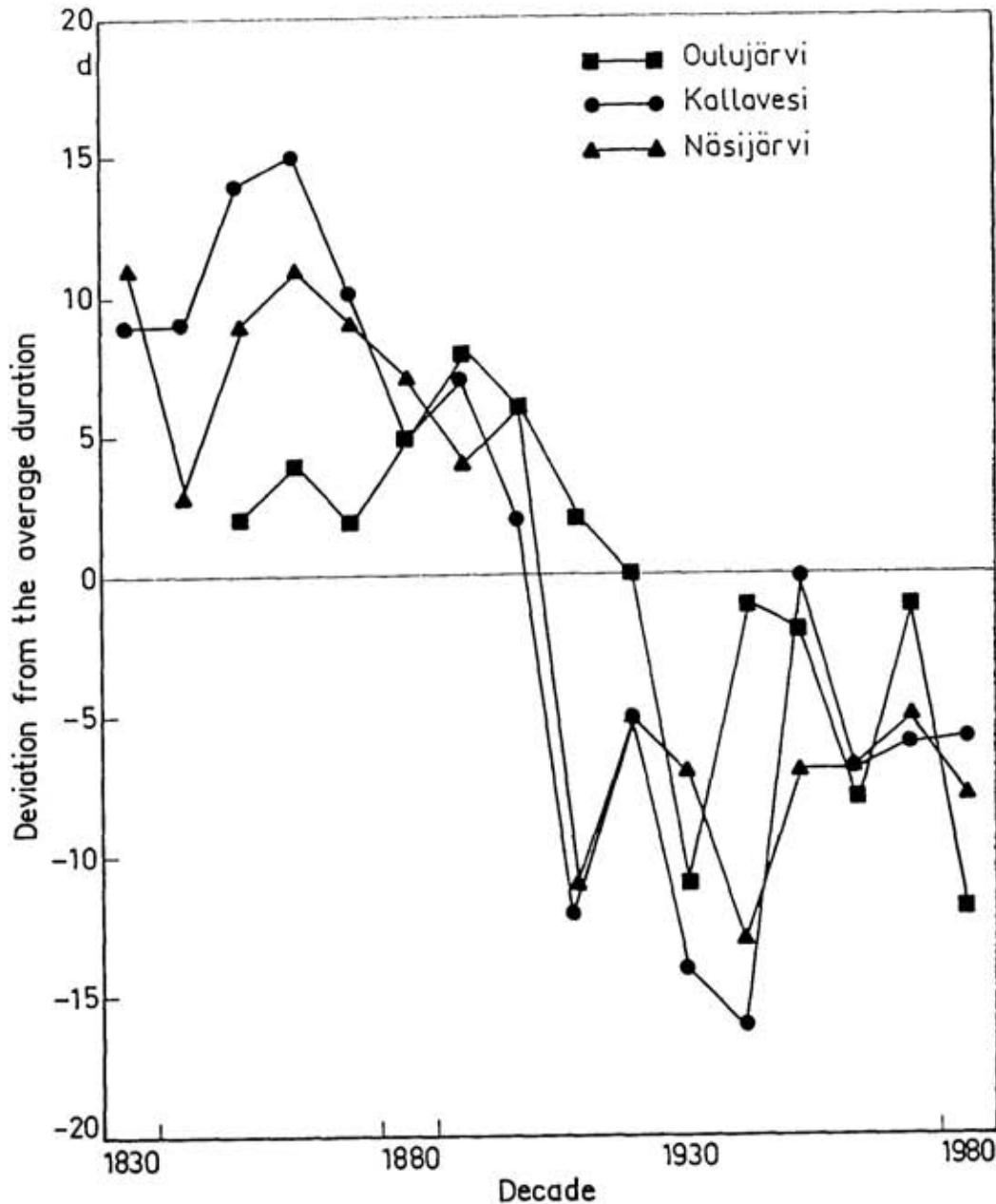
Kuusisto (1987) analyzed the three longest continuous lake ice observation series. The locations of these lakes and the 21-year moving averages of the durations of their ice cover are shown in Fig. 2. The corresponding data from the Gulf of Finland at Helsinki harbour is also included.



**Figure 2.** The 21-year moving averages of the durations of the ice cover on three lakes and in the Gulf of Finland at Helsinki harbour.

Shortening of the ice cover duration from last century until 1940's is evident on all three lakes. A slight lengthening occurred in the 1950's and 1960's. Thereafter, the duration has shown no trend.

The decennial mean durations (Fig. 3) reveal the changes even more clearly than the moving averages. The decennial means were longer than the average of the whole series until the 1900's on Kallavesi and Näsijärvi and until the 1920's on Oulujärvi. After those decades, all the decennial averages have been shorter than the long-term average.



**Figure 3.** The deviations of the decennial average durations of the ice cover from the average of the whole observation period. The first decade is incomplete, on Lake Oulujärvi also the 1890's.

The average and extreme durations of the ice cover on these three lakes have been as follows (in days):

	Average	Max.	Min.
Kallavesi	168	208	96
Näsijärvi	149	188	79
Oulujärvi	189	229	147

The shortest duration occurred on all lakes in the same winter: 1929-1930, when freezing took place very late, as will be discussed later. On the other hand, the longest durations are all from different winters: on Kallavesi from 1866-1867, on Näsijärvi from 1858-1859 and on Oulujärvi from 1880-1881.

The long-term variations of air temperature at least partly correspond to the changes in the duration of the ice cover. The mean temperature in Helsinki became about 1.5°C higher between the 1880's and the 1940's. Slight cooling has occurred since then, but since the year 1988 temperatures have been significantly above average. These temperature changes include the effect of urbanization, which has been of the order of 0.5°C during the last hundred years (Heino 1978).

The connection between climatic variations and the sunspot cycle has recently made a comeback (Friis-Christensen & Lassen 1991). This perpetuum mobile of climatic discussion has now a new uniform; the cycle length is considered as a governing variable instead of the number of sunspots. Fig. 4 shows that a fairly good correlation exists between the length of the sunspot cycle and the duration of ice cover on Lake Kallavesi. The correlation is surprisingly good until the 1930's. With the Helsinki temperatures the length of sunspot cycle correlates well during the whole period of 1860-1990 (Fig. 5).

### Ice Thickness

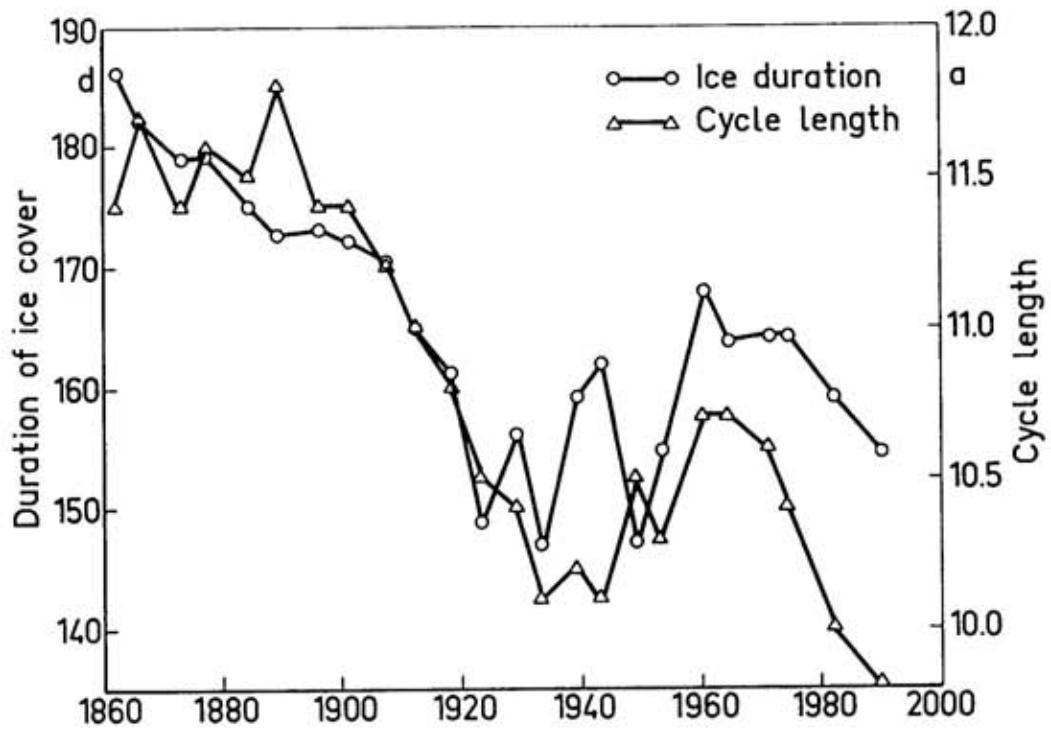
Figure 6 shows the maximum ice thicknesses on three lakes in southern and central Finland in the period 1917-92. There are half a dozen data series of this variable with lengths exceeding 75 years; these three can be considered the most homogeneous of them.

Ice thickness data are sensitive to changes in observation site. On different parts of a lake, the variation of ice thickness may be larger than the average regional variation within a distance of several hundred kilometers. This is partly due to the uneven formation of snow ice (Figs. 7a & b), which also leads to a rather poor correlation between the frost sum and total ice thickness.

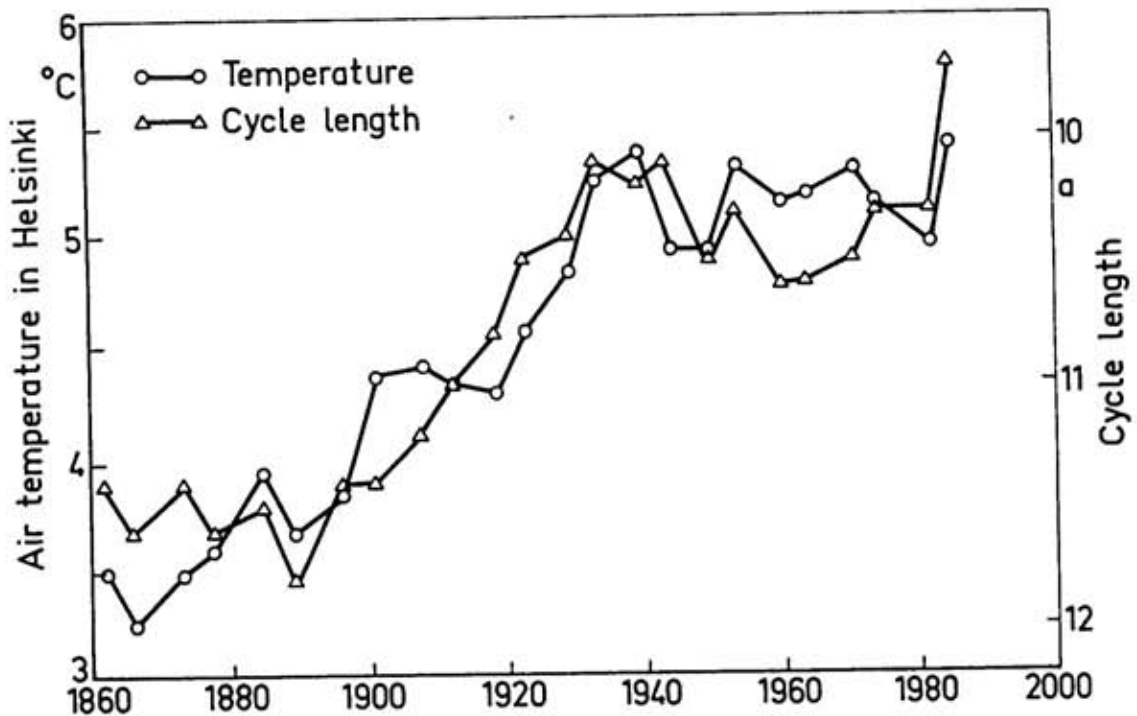
Thus the ice thickness may be of limited value as an indicator of climatic variations. Several meteorological variables - temperature, precipitation, wind velocity and direction - have an effect to ice thickness. The sensitivity of the thickness to morphometric factors further complicates the issue.

### Future Changes

The Finnish Research Program of Climate Change (SILMU) was started in 1991. This extensive five-year program will cover a wide range of topics within meteorology, hydrology, limnology, forestry, agriculture etc. Interim report will be published in June 1992.

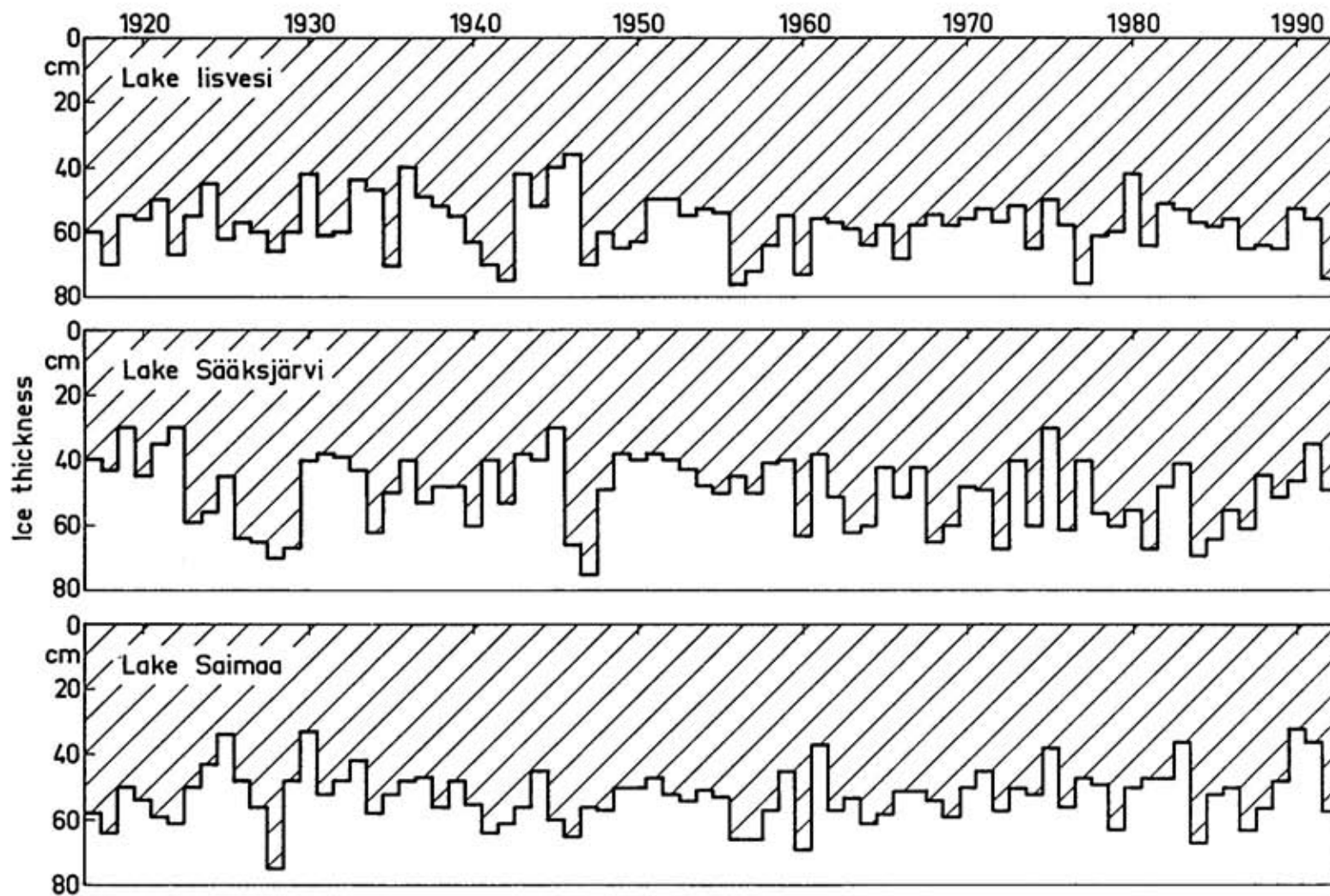


**Figure 4.** The variations of the length of the sunspot cycle and of the duration of the ice cover of Lake Kallavesi. The length of the sunspot cycle is plotted at the central time of the actual cycle, as calculated by Friis-Christensen & Lassen (1991). The duration of the ice cover is averaged for the corresponding periods.

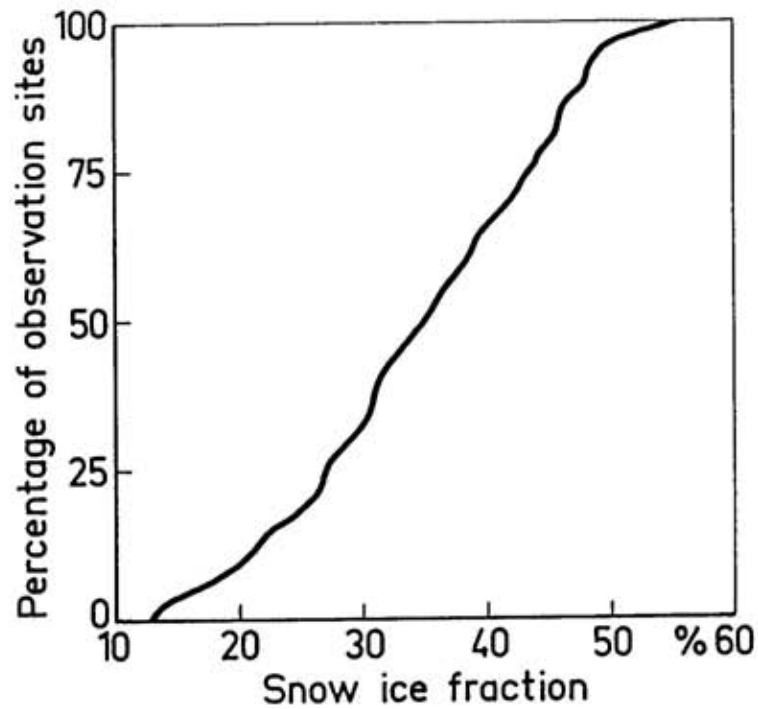


**Figure 5.** The same as Fig. 4, but with the average air temperature in Helsinki.

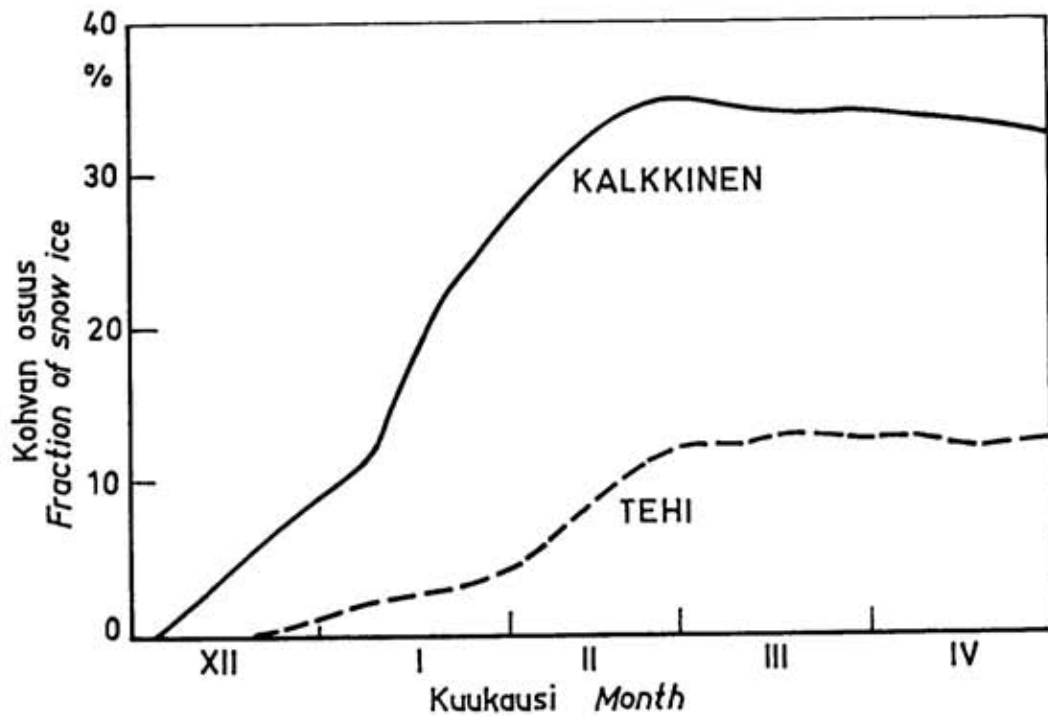




**Figure 6.** The maximum ice thickness on three lakes in southern and central Finland in the period 1917-92.



**Figure 7a.** The fraction of snow ice at the ice thickness observation stations in Finland. The total number of stations is 64.



**Figure 7b.** The development of snow ice fraction at two sites on Lake Päijänne (area 1050 km<sup>2</sup>). The site 'Tehi' represents mid-lake conditions, 'Kalkkinen' is located near the shore.

As to the water sector, the SILMU has the following ongoing research projects:

- Evaluation of changes in hydrological time series
- Microclimatic models for assessing the effect of climate and land cover on evaporation
- The development of soil/plant/atmosphere model
- Effects of climatic change, air pollutants and land use on lake ecosystems
- Effects of the climate change on the hydrography of the Baltic Sea
- The arctic characters of the Northern Baltic pelagial ecosystem
- The effect of climate change on the hydrology and material fluxes of forested catchments
- Effect of climatic change on the temperature of lakes
- Effects of climate change on fishes, fish stocks, fisheries and aquaculture
- Impact of climate change on carbon cycle in freshwater ecosystems

A common scenario of climatic change is recommended for all SILMU projects. It is characterized as a simplified 'best guess' scenario for Finland. The scenario is as follows:

1. Carbon dioxide. The IPCC "Business-as-Usual" projection of atmospheric CO<sub>2</sub> concentration gives values of 350 ppmv in 1990, 430 ppmv in 2020, 540 ppmv in 2050 and 820 ppmv in 2100. Doubling of 1990 levels (to 700 ppmv) occurs in about 2080.
2. Temperature. A linear increase in temperature of 0.4°C per decade, giving temperature increases of 1.2°C by 2020, 2.4°C by 2050 and 4.4°C by 2100. Warming is assumed to be uniform throughout the year and across the whole of Finland.
3. Precipitation. A linear increase in precipitation of 3% per decade, giving increases of 9% by 2020, 18% by 2050 and 33% by 2100. Increases are characterized by enhanced intensities of precipitation events.
4. Unless derived from the above variables (e.g., evaporation), all other climatic variables are fixed at present (baseline) levels. No change is assumed in diurnal, daily or inter-annual temperature variability. The inter-annual variability and within-month distribution of precipitation and the duration of precipitation events are all assumed to be unchanged.

Within the SILMU project, the long time series of freezing, break-up and duration of ice cover will be analyzed. Besides, dynamic models will be applied to selected lakes in order to simulate e.g. their ice conditions in the future.

Both these approaches can help us to understand the sensitivity of lake ice to climatic change. They also form the basis for the estimation of chemical and biological changes in lakes.

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# Monitoring Lake Freeze-Up/Break-Up as a Climatic Index

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## Introduction

There exists a substantial body of literature demonstrating that the formation and melt of lake ice is closely correlated with climatic variables. Extensive investigations of these relationships were carried out more than 50 years ago, for example by Simojoki (1940) for lakes in Finland, a country with innumerable lakes and a substantial network of limnological and meteorological observations. Many subsequent studies in North America and elsewhere have confirmed the main conclusions of the earlier investigators providing additional information on geographical variability of the factors involved. There are countless lakes in northern middle and high latitudes that have a seasonal ice cover; only a few lakes in the High Arctic are permanently ice-covered (Hobbie, 1973, 1984). Nevertheless, the potential value of hemispheric-scale monitoring of lake ice growth and melt (McBeath *et al.*, 1984; Tramoni *et al.*, 1985) as a simple index of seasonal climatic characteristics has not been significantly exploited. This study reviews the findings of studies that bear on this problem and recommends a strategy to utilize the freeze-up/break-up of lakes for climate monitoring.

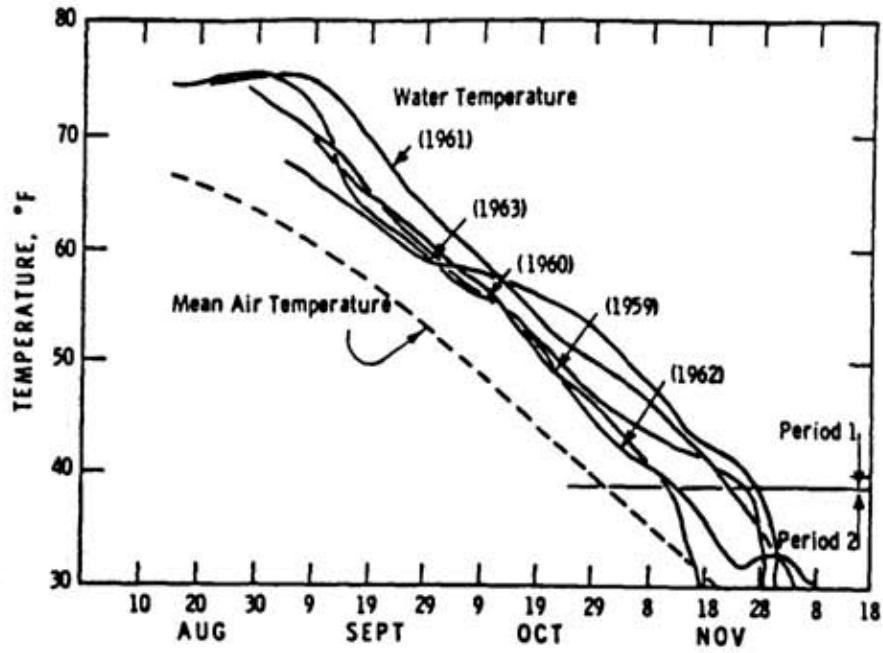
## Criteria for Candidate Climatic Indices

In order for a proxy variable to serve as a climatic index, it should satisfy a number of requirements (Barry, 1984):

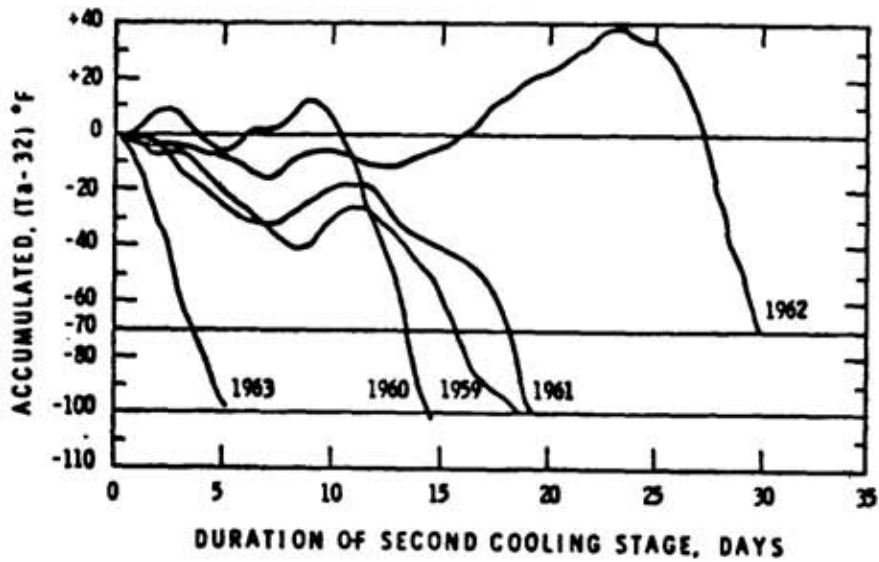
- 1) show high sensitivity to a single meteorological variable;
- 2) have a short time-constant of response;
- 3) have a high signal/noise ratio, at least in some geographical locations that can be pre-selected;
- 4) be readily observable in terms of techniques, frequency of observation, and accessibility of location (in the case of site measurements);
- 5) ideally there should be historical information on the long-term (natural) variability of the selected indicator(s).

## Physical Factors Controlling Freeze-Up/Break-Up

The physical factors affecting lake ice regime are now described. Freeze-up depends on the heat storage in the water (i.e., water depth), the rate and temperature of any inflow, and energy exchange with the atmosphere. For non-flowing water bodies the water cools until the profile becomes isothermal at 4°C (maximum density). Turbulent heat loss formulae are available for this phase (Michael, 1971; Ashton, 1980). The second phase involves cooling of the surface layer until ice nucleation occurs. This takes place in a short interval of time although the process is quite highly weather-sensitive. There is a typical range for Canadian lakes of 2-5 weeks in the date of ice formation about the mean date (Figure 1). For small lakes, empirical relationships between freezing degree-days and the date of ice formation can be useful (Williams, 1971). For Swedish lakes, where winter temperatures are fairly stable and there is considerable snowfall, Bengtsson (1984) finds an increment of 2 cm of ice growth per (freezing degree-day)<sup>0.5</sup>.



a) Water temperature trends, 1959-1963, compared with mean air temperature.



b) Duration of the second cooling stage from 4°C expressed as accumulated freezing degree-days.

**Figure 1.** Variability in freeze-up of McKay Lake (from Williams, 1965).



Simple temperature indices are somewhat less satisfactory for the break-up date, especially where river inflow and wind are important factors, or where there is a high variability in year-to-year snow pack depth on the ice. Predictions of break-up by accumulated thawing degree-days typically give standard deviations of 3-8 days for continental lakes and better results are obtained by using air temperature directly. The late winter ice thickness has some degree of influence on the timing of break-up as suggested in Figure 2, but it is clearly not the sole determinant. Stewart and Haugen (1990) demonstrate for lakes in New York State that lake depth and volume (but not area) are important controls of freeze-up date, in spite of the fact that mean fetch over the lake is expected to be an important control of turbulent heat flux. Morphometry apparently had less effect on break-up dates in New York. For 120 upland lakes in Norway ranging between 0.2 and 6.0 km<sup>2</sup> in area, Skorve and Vincent (1987) find that lake area and the percentage of snow cover in the drainage basin are important independent variables, along with thawing degree-days, in predicting whether a lake remains frozen. However, area is linearly related to the number of streams entering the lake, while percent snow cover is also correlated with degree days. Williams (1965) shows that for a lake near Ottawa, the heat gain at the melting surface over a 10-year period was provided mainly by absorbed short-wave radiation. A complicating factor in areas of heavy snowfall is that the weight of snow depresses the ice causing flooding and the freezing of additional ice on the lake ice cover.

Lake ice formation and decay essentially provides a relatively simple seasonal system that responds on a time-scale of days to surface energy exchanges but can be specified closely in terms of the occurrence of the freeze-up/break-up processes through temperature indices (Michael, 1971). The phase change provides an unambiguous threshold effect and has been already utilized in historical climatology (Simojoki, 1961; Tanaka and Yoshino, 1982). Results for freeze-up and break-up in relation to lakes in northern Canada and Alaska are discussed by Bilello (1980). Figures 3 and 4 illustrate the mean dates of freeze-up and break-up in Canada.

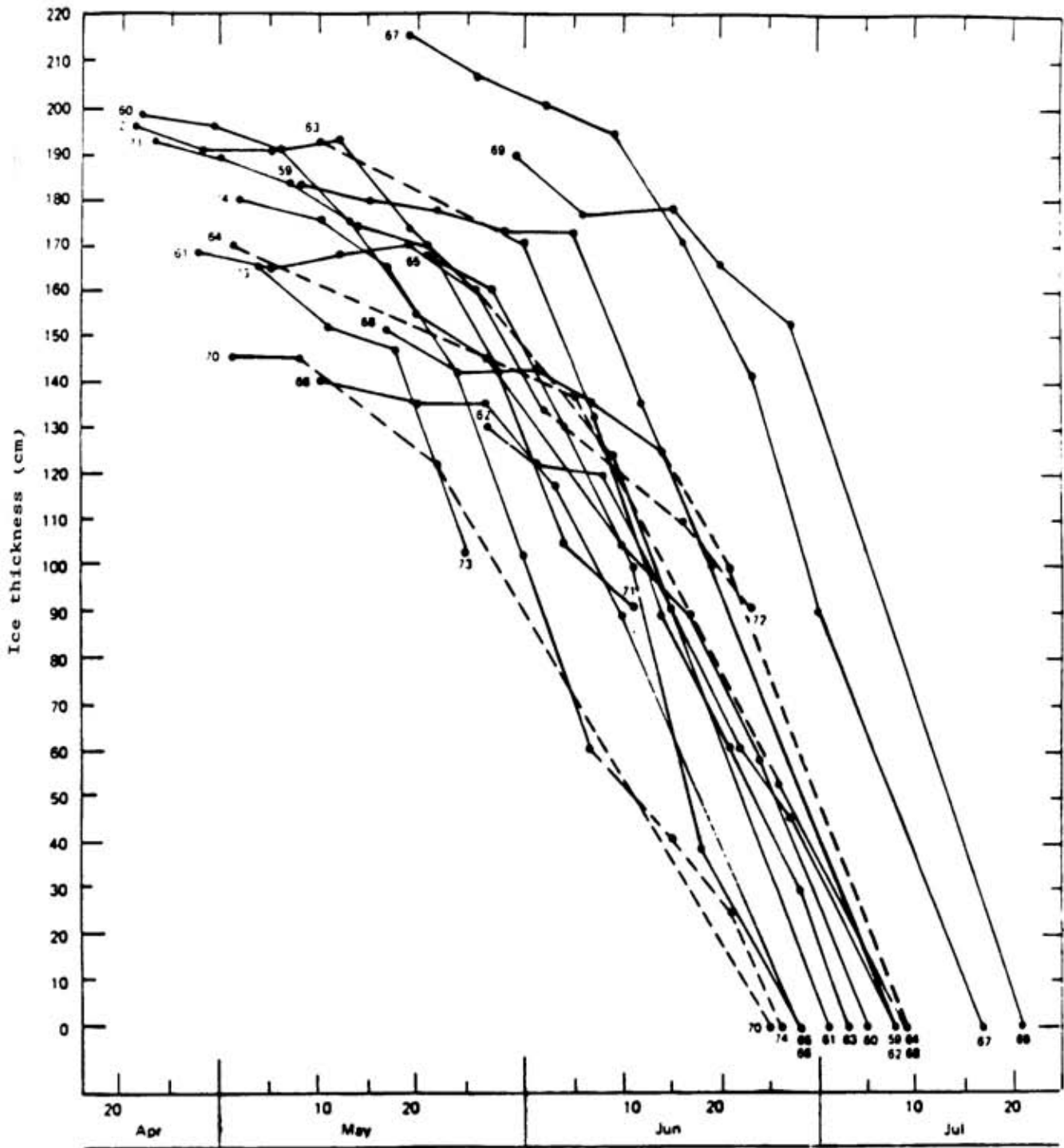
## Observations

Lake ice formation/decay is readily observed on the ground and from the air. It is routinely monitored at more than 300 water bodies across Canada. Historical records are summarized by Burbidge and Lauder (1957), Ragotzkie (1960), Allen and Cudbird (1971) for variable periods beginning in the 1930s/40s. The Canadian records, and subsequent updates, exist in tabular and digital form. They identify the dates of first permanent ice, complete freeze over, first signs of deterioration, the water body clear of ice, and occurrences of incomplete ice coverage/clearance. Analogous work using historical records has been conducted for estuaries (Moodie and Catchpole, 1975). Studies of factors affecting lake ice thickness have also been performed in the Canadian subarctic Adams, 1981; Williams, 1965, 1971). Table 1 lists meteorological stations in Canada that have collected ice thickness data..

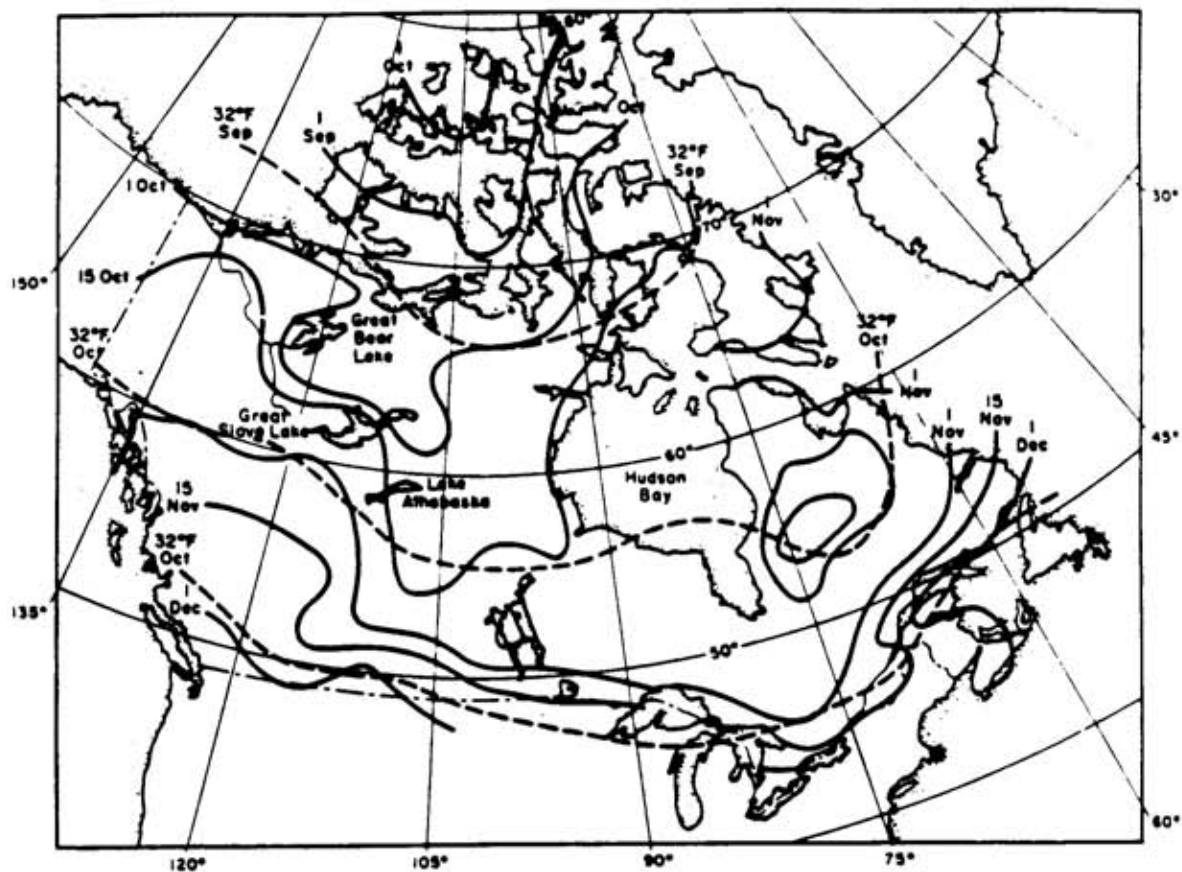
Lake ice data are also collected and reported in Finland (Laasanen, 1982) and Sweden (Meteorologisk-Hydrografiska Anstalt, 1944). Lakes in Finland are mostly shallow; Laasanen (1984) cites mean depths of 3 to 17 m for nineteen lakes. For Canadian lakes, however, such data are mainly unavailable.

## Observational Results

Studies in the literature have focussed either on a few lakes with long records, or on geographical variability for a generally shorter time interval. For example, Ruosteenoja (1986) developed regression models for the break-up on Lake Kallavesi, a large lake, in



**Figure 2.** Illustration of the effect of ice thickness on break-up date for Ennadai Lake, N.W.T. (from Bilello, 1980).



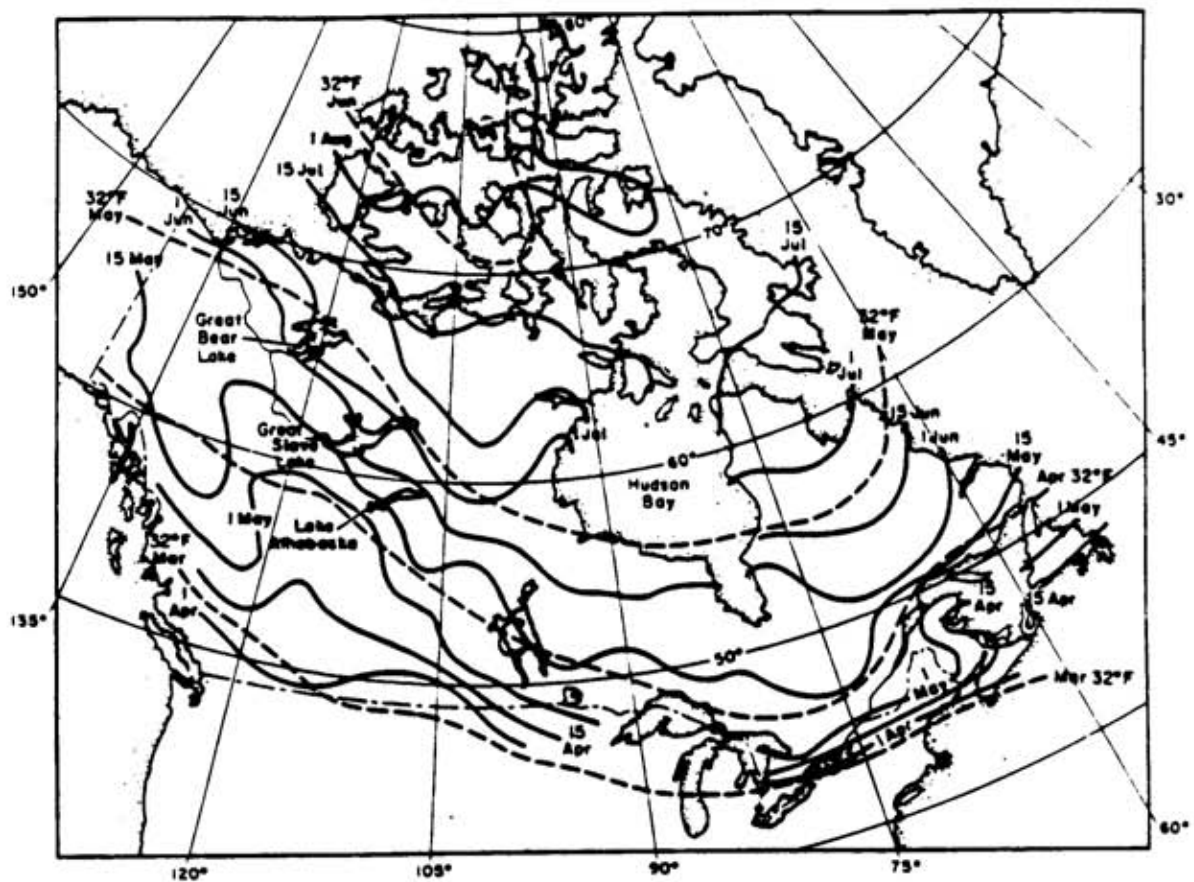
— Isoleths at half month intervals give the average dates of freeze-up on rivers and lakes.  
 - - - 32 degrees Fahrenheit isotherm for month shown.

**NAVIGATION SEASON**

Lake or bay	Closes
Great Bear Lake	15 Oct-1 Nov
Great Slave Lake	15 Oct
Lake Athabaska	20 Oct
Hudson Bay	Mid Oct*

\*Covered with ice floes by Jan.

**Figure 3.** Mean date of freeze-up of freshwater bodies in Canada (from Allen and Cudbird, 1971).



— Isoleths at half month intervals give the average dates of breakup of rivers, and the appearance of breaks along the shore of lakes.  
 - - - 32 degree Fahrenheit isotherm for month shown.

**NAVIGATION SEASON**

Lake or bay	Open	Ice remains until
Great Bear Lake	Mid July	August
Great Slave Lake	15 June	Late July
Lake Athabasca	1st wk. June	
Hudson Bay	Mid July	August

Figure 4. Mean date of break-up of freshwater bodies in Canada (from Allen and Cudbird, 1971).

**Table 1.** List of Canadian stations

<i>Canadian stations</i>	<i>Body of water</i>	<i>Years of record</i>	<i>Fresh or salt water</i>
Alert	Dumbell Lake	1959-1974	F
Alert	Parr Inlet	1959-1974	S
Arctic Bay	Arctic Bay	1960-1969	S
Baker Lake	Baker Lake	1959-1974	F
Brochet	Reindeer Lake	1959-1974*	F
Cambridge Bay	Cambridge Bay	1959-1974	S
Cape Dorset	Cape Dorset Harbour	1971-1974	S
Cape Parry	Amundsen Gulf	1960-1974	S
Cartwright	Cartwright Harbour	1960-1974*	S
Chesterfield Inlet	Hudson Bay	1960-1974	S
Churchill	Churchill River	1960-1973*	Ft
Clyde River	Patricia Bay	1960-1974	S
Coppermine	Coppermine River	1959-1974	F
Coral Harbour	Munn Bay	1959-1974	S
Cree Lake	Cable Bay in Cree Lake	1971-1974	F
Ennadai Lake	Ennadai Lake	1959-1974	F
Eureka	Slidre Fiord	1959-1974	St
Ft Chipewyan	Lake Athabasca	1962-1974	F
Frobisher Bay	Koojesse Inlet	1959-1974	S
Goose Bay	Terrington Basin	1959-1974	Ft
Hall Beach	Foxe Basin	1959-1974	S
Holman Island	Kings Bay	1960-1968	S
Hopedale	Hopedale Harbour	1961-1974*	S
Inoucdjouac	Innuksuak River	1959-1974*	Ft
Inuvik	Mackenzie River	1961-1974	F
Isachsen	Deer and Louise Bays	1959-1974*	S
Island Lake	Island Lake	1971-1974	F
Moosonee	Moose River	1960-1974	Ft
Mould Bay	Mould Bay	1959-1974	S
Nitchequon	Lake Nichicun	1959-1974	F
Norman Wells	Mackenzie River	1959-1974*	F
Norway House	Nelson River	1957-1974*	F
Pond Inlet	Eclipse Sound	1965-1970*	S
Resolute	Resolute Bay	1959-1974	S
Sachs Harbour	Amundsen Gulf	1959-1974	S
Schefferville	Knob Lake	1961-1974	F
Schefferville	Maryjo Lake	1961-1968	F
Spence Bay	Spence Bay Harbour	1960-1968*	S
Trout Lake	Big Trout Lake	1961-1974	F
Yellowknife	Back Bay	1959-1974	F

\*Some years for the period of record shown are missing

†These bodies of water may be brackish

central Finland, extending earlier work of Simojoki (1961). Robertson (1988) analyzed the ice and temperature records since 1856 for Lake Mendota (3,940 ha) at Madison, Wisconsin. Palecki and Barry (1986) used 63 lake records from Finland to develop regression models of spring or fall temperatures in relation to dates of final break-up and final freeze-up, respectively. Correlations for southern Finland in November were 0.88 and in April 0.69 between a composite of lake ice dates and a single meteorological station. Air temperature anomalies for November and April calculated from a regression based on freeze-up and break-up dates of Lake Kallavesi compared well with observed temperatures at Helsinki using 20-year moving averages, for example. Other recent studies of ice cover on Finnish lakes have been carried out by Laasanen (1984) and Kuusisto (1989).

### **Climate Change Detection**

The dates of lake freeze-up/break-up have been correlated with various climatic and other variables. Essentially two types of correlation have been made: either using data on a climatic variable such as air temperature for a fixed calendar period, or using a moving index of freezing, or thawing, degree-days accumulated up to the date of the specified event. Robertson (1988) shows that by either method similar changes in date of freeze-up or break-up are obtained per 1°C change in temperature. Numerous results appear to indicate that in the middle latitudes of the Northern Hemisphere, a 1.0°C change in autumn or spring air temperature gives rise approximately to a 4-6 day change in mean freeze-up date or break-up date, respectively (Palecki and Barry, 1986; Robertson, 1990). The advantage of lake freeze-up/break-up as a climatic index is that a single date is highly correlated with an integrated temperature record for the transition season months, as illustrated in Figure 5 for southern Finland.

### **Remote Sensing of Ice Formation and Break-up**

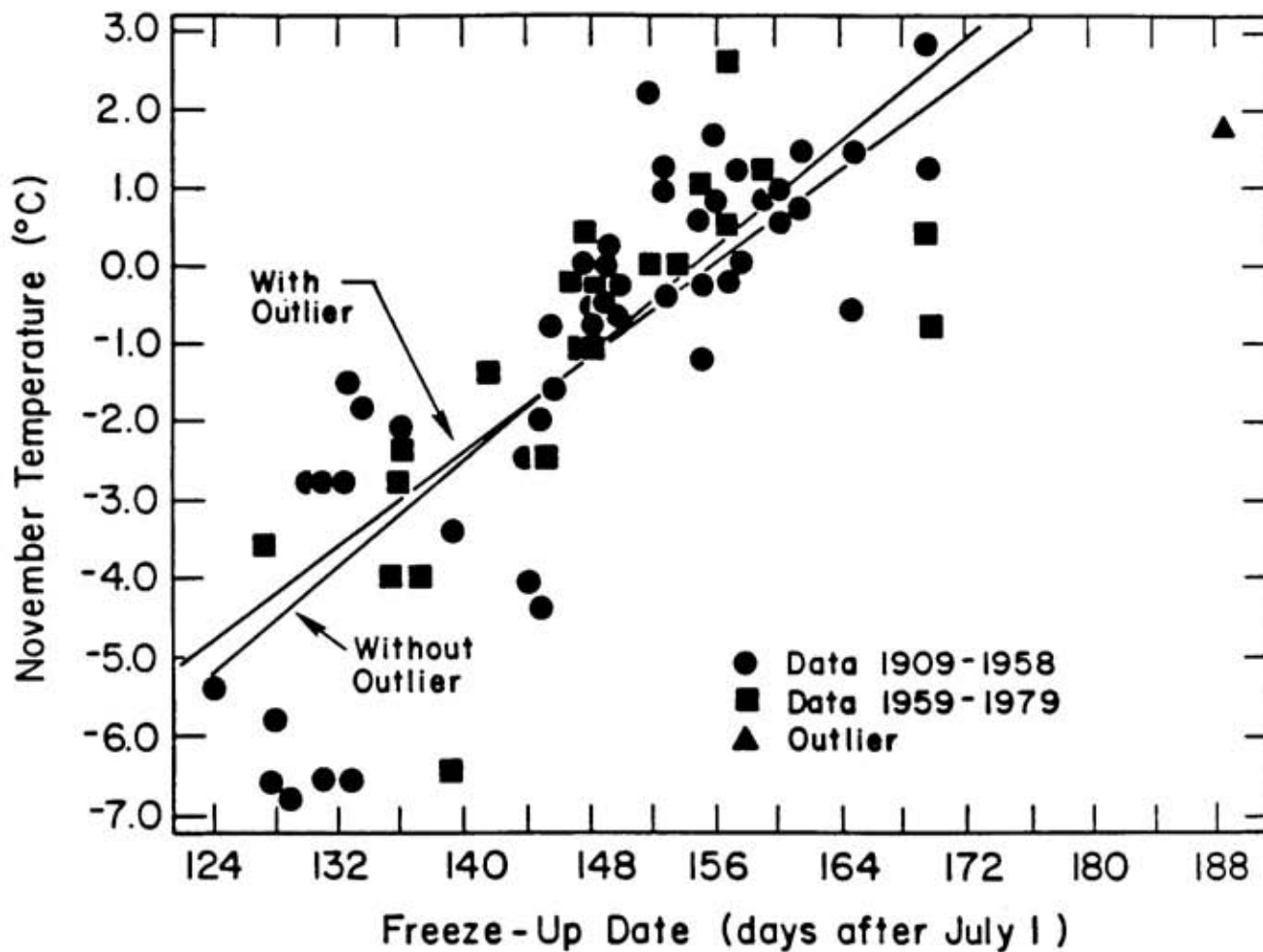
The global monitoring of lake ice conditions can be carried out routinely through satellite remote sensing. This can provide regular, consistent observations using a uniform approach. The ability to detect the timing of ice formation/break-up on lakes depends on:

- 1) spectral separation of lake ice from other features;
- 2) consistent, identifiable criteria of ice formation/break-up.
- 3) the availability of cloud free conditions over the lake, if visible band data are to be used
- 4) a sufficiently large lake to avoid shoreline contamination of the spectral signature in the data being processed;
- 5) sufficiently frequent coverage to provide precision in the date of an identified change.

Of these requirements, perhaps the most significant is the need for frequent, typically daily, observations in order to detect freeze-up and break-up as precisely as possible. The only practical remote sensing platforms that offer such frequent coverage are the meteorological satellites such as the NOAA, DMSP, and Meteosat series. The tradeoff is the lower spatial and spectral resolution offered by these sensors compared to Landsat and SPOT, for example. However, data and processing costs are much lower - to the extent where monitoring on a global scale becomes feasible.

Spectral signatures of various snow and lake ice-melt states are reported by Bolsenga (1983). Dean and Ahlnaes (1984) and Maslanik and Barry (1987) confirm that these





**Figure 5.** Regression of November temperature at Jyväskylä and lake freeze-up in southern Finland for 1909-59 and 1959-79 (from Palecki and Barry, 1986).

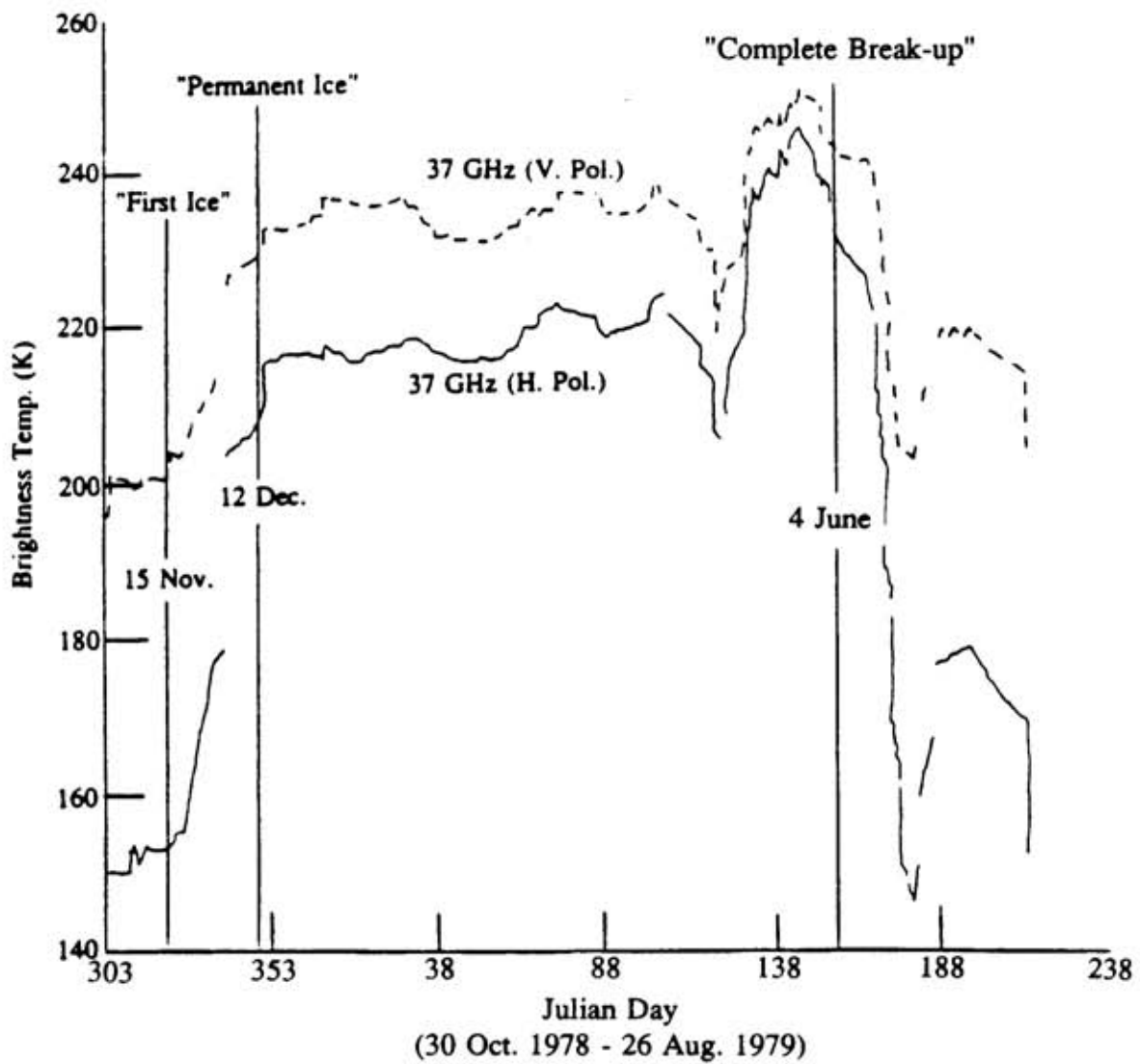
spectral properties enable satellite remote sensing of lake freeze-up/break-up to be utilized although care is required in distinguishing new and sediment-loaded water ice from open water (Maslanik and Barry 1987). Comparisons of the spectral response curves of different sensors (DMSP Operational Line-scan System [OLS], Landsat Thematic Mapper [TM], and NOAA Advanced Very High Resolution Radiometer [AVHRR]) relative to the reflectance of different water and ice conditions point out the value of spectral channels in the blue and green wavelengths to discriminate between water and ice (Maslanik and Barry, 1987). In particular, detection of freeze-up proved difficult using only the broad-band visible channel on OLS. Whereas freeze-up appears as a gradual change in reflectance until snowfall covers the ice, break-up generally manifests itself as a clearly-observable rapid drop in reflectance as the ice cover deteriorates. However, as noted earlier, break-up dates are affected by several variables in addition to air temperatures, and thus are more problematic than freeze-up dates as a proxy data set for climatic change. It is also critical that a consistent definition of what constitutes "freeze-up" and "break-up" in the imagery be defined. When comparing dates determined using remotely-sensed data to ground observations, it is also important to recognize that ice-out conditions as visible from a shore location may not be directly comparable to conditions viewed from orbit.

A systematic relationship was found between lake break-up dates in Canada and Finland as observed with DMSP OLS visible-channel images and reported by ground observers. Detection of freeze-up was hindered, however, by more cloudiness in autumn in the areas studied. Skorve and Vincent (1987) used principally Landsat visible band 0.6-0.7  $\mu\text{m}$ , enlarged to 1:250,000 scale, to study ice melt on upland lakes in southwestern Norway.

Microwave data can eliminate the problem of cloud-covered scenes. Passive microwave currently offers only low resolution (ca. 25 km), thereby limiting its use to lakes of  $\geq 2,500 \text{ km}^2$  (Hall *et al.*, 1978, 1981). However, the potential of passive microwave data is worth exploring further. A case study using an annual time series of Nimbus 7 Scanning Multichannel Microwave Radiometer (SMMR) data for Great Slave Lake in Canada (Figure 6) shows that the inflection points in the data agree quite well with the ice status reported by ground observers. "Complete Breakup" conditions as seen from shore precede the date indicated by SMMR by about 20 days, which suggests that the SMMR data are detecting ice within the central part of the lake, since ice-out conditions begin near shore first. Synthetic aperture radar (SAR) data can provide up to 10m resolution but there is a risk of insufficiently frequent coverage due to the swath width and the variable interval between repeat orbits. Also, little difference may exist between the backscatter of smooth ice and open water, except for lakes frozen solid to the lake bed.

### **Concluding Remarks**

In summary, lake ice freeze-up and break-up dates have been shown to have significant statistical relationships to surface air temperatures, at least over local to regional scales. Use of these dates for detection of climatic change on a global scale will require sampling that, in practical terms, can only be done using remote sensing methods. Digital data from meteorological satellites offer the best hope for a systematic sampling program, although problems exist due to cloud cover, quantitative definition of freeze-up and break-up appropriate to remote sensing, as well as the spatial and spectral resolution of the sensors themselves. Satellite sensors planned as part of the Earth Observing System offer improved spatial and spectral resolution, and selection of "indicator lakes" with adequate size and shape would help remove some of the uncertainty of a remote sensing program for large-scale applications.



**Figure 6.** Time series of Nimbus 7 Scanning Multichannel Microwave Radiometer (SSMR) brightness temperatures for Great Slave Lake, Canada, for 30 Oct. 1978 - 26 Aug. 1979 with dates of changes in ice status as reported by surface observers. Open water in the SMMR data is indicated by low brightness temperatures and large polarization differences.

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# Active and Passive Microwave Remote Sensing of Frozen Lakes for Regional Climate Studies

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## Introduction

Lake ice development and dissipation is influenced by meteorological factors, such as air temperature and snow depth, as well as by limnological factors such as water temperature and impurity content. Thus lakes can be sensitive indicators of regional climate, especially if measured over a period of many decades (Rist, 1970; Palecki et al., 1985; Palecki and Barry, 1986).

Lake ice, even if it is actively melting, can be measured easily on visible and near infrared satellite imagery after the surrounding snow has melted. The resolution of the Landsat multispectral scanner (MSS) (80 m) or thematic mapper (TM) (30 m) sensors, respectively, is adequate for analysis of lake ice formation and dissipation on many lakes. However, use of data acquired in the reflective part of the electromagnetic spectrum is limited because the lakes can only be monitored during the daytime and under cloud-free conditions. In addition, the repeat cycle (16 or 18 days) of the Landsat satellites is inadequate for reliable determination of lake ice freeze-up and break-up dates. Furthermore, even though Landsat data were acquired beginning in 1972 and therefore should provide about 20 years of data, some of the early Landsat data are no longer available in digital form. Beginning in the late 1960s, NOAA sensors have acquired data with a 1.1 km spatial resolution and daily coverage that is suitable for lake ice monitoring. However, these data are suitable only for the study of large lakes.

Aircraft and satellite-borne Synthetic Aperture Radar (SAR) data have also been employed for studying lakes and are particularly useful for studying thaw lakes in northern Alaska (Sellmann et al., 1975a; Weeks et al., 1977; Mellor, 1982). Together, Landsat and SAR data have been used to determine categories of ice thickness and thus to infer lake depth (Sellmann et al., 1975b). SAR sensors are advantageous because data can be acquired in darkness, and through cloudcover. Additionally, the radar backscatter is related to internal structure of lake ice, and the resolution of the SAR data is suitable for monitoring of lakes having a range of sizes.

Passive microwave data are useful for studies of large lakes. The microwave brightness temperature, a measure of naturally-emitted microwave energy which is a function of emissivity and temperature, is related to ice thickness at certain microwave wavelengths. Passive microwave data of frozen lakes can usually be acquired through cloudcover and darkness. In this paper, active and passive microwave remote sensing techniques and their potential for analysis of lake ice are discussed and reviewed.

## Lake Ice Freeze-up and Break-up and Climate

Lake ice break-up date is a key parameter to measure for analysis of regional climate and can potentially be monitored via satellite. Break-up can be defined in several ways but

it is important to ensure that the same 'event' is measured each year. For example, Williams (1971) defines break-up as the date that a lake is completely free of ice.

Long-term break-up records in North America show that from about 1870-1940, break-up of lakes occurred at progressively later dates; however between 1940 and 1971, no apparent change in break-up date was observed (Williams, 1971). Schindler et al. (1990) have shown that air and lake temperatures in the Experimental Lakes Area (ELA) in northwestern Ontario, have increased by 2<sup>o</sup> C and the length of the ice-free season has increased by 3 weeks according to 20 years of records. However, no change in freeze-up date was observed. They attribute the change observed in the ELA to air temperature increases in March, April and May and below average snow covers in March, resulting in less snow late in the season to insulate lake ice.

The precise date of freeze-up depends mostly on wind and cloud cover and the rate of freezing, and maximum ice thickness is dependent upon air temperature conditions and snow cover according to Hobbie (1973).

Deterioration of lake ice is dependent upon amount of solar radiation, as well as on thickness and type of ice and on wind conditions and air temperature. A few centimeters of snow effectively block penetration of light into lakes; when snow melts, solar heating may begin (Hobbie, 1973; Schindler et al., 1990). Similarly, the 'white ice' that often forms from refrozen snow on the surface of lakes takes longer to melt than does clear ice that forms from the lake water because the white ice is less effective in absorbing incident radiation (Hobbie, 1973; Adams, 1976). Thus the amount of white ice is also important in determining the rate of break-up.

### **Synthetic Aperture Radar**

The resolution of the satellite-borne SARs is suitable for the study of lakes of a range of sizes. Data from the L-band (23.5 cm wavelength) Seasat SAR, which operated between June and October 1978, has a spatial resolution of approximately 25 m, and the C-band ERS-1 SAR, currently in operation, has a resolution of approximately 30 m. Furthermore, since radars at these frequencies are unhindered by cloudcover and darkness, useful data should be acquired during every satellite overpass (Hall and Ormsby, 1983).

Side looking airborne radar (SLAR) data have been obtained of frozen lakes since the early 1970s. For example, Bryan and Larson (1975) classified freshwater lake ice into several distinct categories using SLAR X- and L-band data in Whitefish Bay and the Straits of Mackinac, Michigan. More recently, Leconte and Klassen (1991) found that ice that formed in the western part of Split Lake in northern Manitoba gave strong SAR returns. The strong backscatter appeared to be caused by refrozen slush that formed between the lake ice and snow ice during the initial stages of ice formation.

Sellmann et al (1975a) used both Landsat and SLAR data to classify lakes in northern Alaska and to determine whether the ice cover on the lakes was frozen to the lake bed or not. During the 1970s and early 1980s, the thaw lakes located in northern Alaska were studied extensively using airborne radars (Sellmann et al., 1975a and b; Elachi et al., 1976; Weeks et al., 1977, 1978 and 1981; Mellor, 1982). These lakes are formed by the thawing of ground ice in continuous permafrost, and are oriented in a northwest-southeast direction that is approximately perpendicular to the direction of the prevailing winds.

At approximately the time of maximum ice thickness (March through early May), relative ice thickness and thus lake depth can be determined using both SAR and Landsat

data. By noting the relative melt dates of the ice on many of the thaw lakes, Sellmann et al. (1975b) were able to classify the lakes into 3 depth categories. The deepest lakes, which contain the thickest ice, retain their ice covers longer than do shallower lakes containing thinner ice. (This is, of course, because the thinner ice melts sooner than does the thicker ice.) Two different categories of lakes give distinctly different radar responses: lakes that are frozen to the bottom give low (dark) returns, and lakes in which water underlies an ice cover give high (bright) returns (Sellmann et al., 1975a). Thus a lake that retains its ice cover into July or even August, as determined from the Landsat data, and gives a high radar return, is deeper than about 2 m. This categorization is possible because lakes that are not frozen to their beds (ie., where water is present between the ice cover and the lake bed) are more reflective to the radar signal than are lakes that are frozen to their beds. This response has been observed both using X- and L- band radars in northern Alaska. This ability to distinguish floating ice from ice frozen to the lake bed does not necessarily apply in other geographic areas.

An explanation for the different radar returns for lakes that are frozen to their beds and lakes that contain floating ice in northern Alaska has been offered. The dielectric constant of freshwater ice and water is 3.2 and 80, respectively while the dielectric constant of frozen sediment is approximately 4 at the L-band frequencies. Stronger radar reflections occur at interfaces with a greater difference in dielectric constant (Elachi et al., 1976). Thus lake ice that has water below will reflect a radar signal more than lake ice frozen to the bed. However, the dielectric discontinuity between the ice and water alone cannot explain the SAR returns from the Alaskan North Slope lakes because at the non-perpendicular angles at which the radar beam encounters the ice, and in lieu of other scatterers, a radar signal is reflected away from the sensor even in the presence of a highly reflective ice/water interface. Weeks et al. (1978 and 1981) and Mellor (1982) have shown that the numerous elongated or columnar bubbles that are present in the ice contribute to the high returns from the floating ice. These bubbles are oriented perpendicular to the forward scattering of the radar beam (toward the ice/water interface or ice/sediment interface). Bubbles with this orientation and adequate size relative to the radar wavelength cause additional scattering in lakes that are not frozen to their beds, and the scattering is directed toward the sensor and contributes to the observed high SAR returns (Weeks et al., 1978 and 1981).

Lakes or parts of lakes that are deeper than about 4 m in northern Alaska have few columnar bubbles as compared to shallower lakes, and do not give high radar returns even though they contain floating ice. Few bubbles will form when there is a substantial amount of water beneath the ice to accept the gases being emitted without the water becoming saturated (Mellor, 1982). The central and western parts of Teshekpuk Lake, the largest of the thaw lakes in northern Alaska, are considerably deeper than the rest of the lake, and though the ice was not frozen to the bottom, the radar backscatter (X band in this case) was found to be low by Mellor (1982). Thus, to give the high radar returns, the lakes must be generally shallow (less than about 4 m deep), have numerous columnar air bubbles, and not be frozen to the bottom. Many lakes on a world-wide basis do not fulfill all of these requirements and thus do not provide the aforementioned radar returns that are characteristic of the North Slope lakes. Future study of SAR data of the North Slope and other lakes should reveal other interesting SAR signatures that relate to freeze-up, ice thickness, bubble structure and break-up.

### **Passive Microwave Remote Sensing**

Passive microwave data have a considerable potential for lake ice studies, particularly for the study of large lakes. The microwave brightness temperature is directly related to ice thickness (Swift et al., 1980a and b; Hall et al., 1981) and the detection of passive



microwave energy is usually unaffected by cloudcover in areas where the clouds contain little or no liquid water which is quite often the case over frozen lakes. The daily coverage provided by the Scanning Multichannel Microwave Radiometer (SMMR) which was flown on the Nimbus-7 satellite and operated from 1978-1988, and the Defense Meteorological Satellite Program (DMSP) Special Sensor Microwave Imager (SSMI), is excellent for lake ice studies. In addition, the acquisition of passive microwave data is unhindered by darkness. The major drawback to the use of passive microwave satellite data is the coarse spatial resolution (usually  $> 15 \text{ km}^2$ ) and thus passive microwave satellite data are currently only useful for the study of large lakes. Also, the longer microwave wavelengths are more useful for lake ice thickness studies, and these are not currently available on satellite-borne passive microwave sensors.

A uniform slab of freshwater ice will emit microwave radiation in a quantity proportional to its thickness. In clear ice where there are few scatterers, the emissivity is affected primarily by the ice/air interface, the ice/water interface, and the ice thickness. The numerous air/ice interfaces (bubbles) within the ice sheet, and the ice/water interface at the bottom of the ice sheet scatter the microwave radiation and lower the emissivity. In lake ice there can be numerous bubbles that may be distributed unevenly throughout the ice causing non-uniform brightness temperatures across a lake. For smooth surfaces, the emissivity of freshwater ice depends upon the attenuation coefficient (a function of temperature) and the mean thickness of the ice (Swift et al., 1980a and b).

Several studies have demonstrated the value of passive microwave remote sensing for analysis of freshwater lake ice. Brightness temperature will decrease with increasing wavelength (or increasing penetration through the ice) (Schmugge et al., 1974; Hall et al., 1981). And the longer microwave wavelengths appear to be potentially useful for ice thickness determination.

Longer wavelength microwave radiation can be sensed from deeper within the ice than can shorter wavelength radiation. For freshwater lake ice studies, shorter microwave wavelengths, e.g. 0.81 cm (37 GHz) to 1.4 cm (22.22 GHz) sense the snow overlying the lake ice while longer wavelengths sense the entire thickness of the ice, and in general, the ice/water interface (Schmugge et al., 1974; Hall et al., 1981). At the shorter wavelengths, the overlying snow is contributing more to the observed brightness temperature than is the ice because snow grains are large enough with respect to the wavelength of the radiation to cause scattering. However, snow grains are generally not large enough, relative to 6.0 or 21.0 cm wavelength radiation to cause significant scattering; thin, dry snow overlying fresh water ice is almost transparent to the microwave radiation emanating from the ice below.

Swift et al. (1980a and b) flew a 6.0 cm aircraft-borne radiometer over the Mackinac Straits of northern Michigan simultaneous with the acquisition of in-situ ice thickness measurements. The ice was 60-80 cm thick and brightness temperatures ranged from 200-220 K. Another 6.0 cm radiometer was flown over Walden Reservoir, Colorado when the lake ice was 65-70 cm thick and resulting brightness temperatures were also 200-220 K (Hall et al., 1981). Additionally, Hall et al. (1981) found a good correlation between brightness temperature and ice thickness on 6 flights over Walden Reservoir between 1977 and 1980. The longer wavelengths provided better correlations between ice thickness and microwave brightness temperature than did the shorter wavelengths.

Studies of Great Bear and Great Slave lakes in northern Canada have been conducted by Walker and Barry (this volume) using SSMI data to detect and monitor the formation and development of ice during the 1991-92 winter using the 85 GHz channel data. The frequencies available with the SSMI (85.5, 37.0, 22.235 and 19.35 GHz) are not well-

suited to studying lake ice thickness, but are suitable for monitoring the development and dissipation of lake ice. Longer wavelengths, to be available on future sensors, will be better suited to monitor lake ice thickness.

## Conclusions

Active and passive microwave data, especially if combined with data from the reflective part of the spectrum, provide the potential for analysis of key attributes of lake ice that are climatologically significant. Lake ice thickness can be assessed at least in a relative way, using both active and passive microwave remote sensing. Whether or not a lake is frozen to its bed can be assessed under some circumstances. Information concerning the freeze-up and break-up processes can also be assessed.

Because lakes can freeze or melt quickly, it is important to have frequent observations for monitoring purposes. Thus the use of the microwave part of the spectrum is ideally suited because cloud cover generally does not preclude the acquisition of data, and data can be acquired during darkness. Additionally, the microwave energy emanates from the entire thickness of the ice cover and can thus provide information on the internal structure of the lake ice which may be related to ice thickness because of the bubble structure. Passive microwave satellite data, usually acquired on almost a daily basis, is suitable for the study of large lakes, while SAR data, though acquired less frequently, is suitable for more detailed studies of both small and large lakes.

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# Snow Modelling in General Circulation Models (GCMs)

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General circulation models are mathematical models of the atmosphere based on the physical conservation laws for momentum, energy and mass (chiefly water vapour). For a fluid on a rotating body subject to an external heating source, these take the form of nonlinear partial differential equations, which cannot be solved analytically. Solutions are therefore obtained by numerical means. The atmosphere is subdivided into several vertical layers, with transfers between them being calculated using finite difference or finite element formulations; horizontal transfers within each layer are calculated at discrete grid points over the earth using finite difference techniques (in the case of grid-point models) or truncated expansions of spherical harmonics (in the case of spectral models). Thus, the state of the atmosphere at a given point in time is represented by values of temperature, humidity, wind velocity etc. at each grid point for each level over the globe; the governing equations are integrated forward in discrete time steps from a set of specified initial conditions.

The "physics" portions of the calculations (dealing with small-scale motions and thermodynamics) involve highly complex calculations and parametrizations of sub-grid scale processes, and are therefore treated separately from the "dynamics" portions (dealing with large-scale atmospheric motions). In spectral models as well as in grid-point models, the equations describing the "physics" are solved in normal three-dimensional grid space and the resultant time rates of change of the atmospheric variables are added to the rates of change obtained from the dynamics. The processes treated in this manner include gravity-wave drag and turbulent diffusion, cumulus convection, cloud formation and precipitation, radiation absorption by the atmosphere, surface transfers of energy and momentum, and land surface hydrology.

The modelling of snow in GCMs is closely related to the treatment of the soil profile in the land surface schemes they employ. In general, soil temperatures are calculated by making use of one of two basic approaches. The first involves the so-called force-restore method of Deardorff (1978), which assumes diurnally periodic surface forcing and obtains the temperature of a vanishingly thin surface layer on the basis of a forcing term depending on the ground heat flux at the surface and a restoring term depending on the underlying, bulk soil temperature. The second involves dividing the soil profile into two to four layers and calculating heat fluxes between them based on finite differencing methods. The latter strategy has the advantage of being able to realistically handle annual as well as diurnal forcings. In almost all cases, however, snow is thermally lumped together with the ground; only one land surface scheme currently in use in a GCM (Verseghy et al., 1992; Verseghy, 1991) models snow cover as a separate thermal layer. For the most part, snow is considered as affecting only the albedo and heat capacity of the soil.

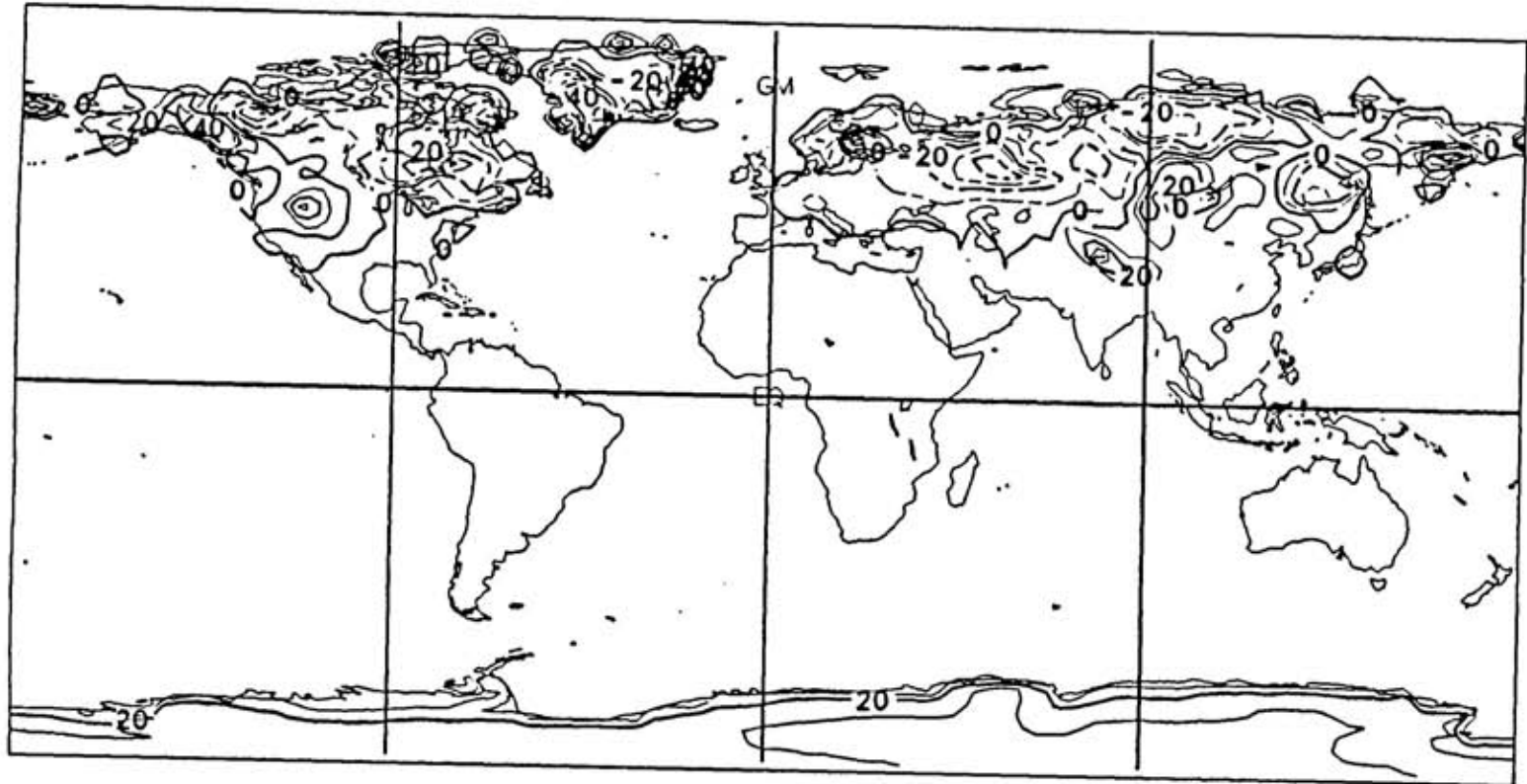
Two basic approaches are likewise used to model the soil moisture regime. The first makes use of the so-called bucket model, introduced by Manabe (1969), in which the soil is treated as a single-layer reservoir with inputs and outputs at the surface only (via evaporation and infiltration of precipitation). This method has the advantages of simplicity

and cheapness, and is still widely used. However, as will be shown below, it introduces unrealistic lags into the surface temperature during phase changes in the soil. The other strategy, as in the modelling of soil temperatures, is to divide the soil profile into two to four layers and calculate the water fluxes at the layer interfaces based on accepted hydrological theory. In these models too, however, snow is usually treated in effect like a second frozen soil water store.

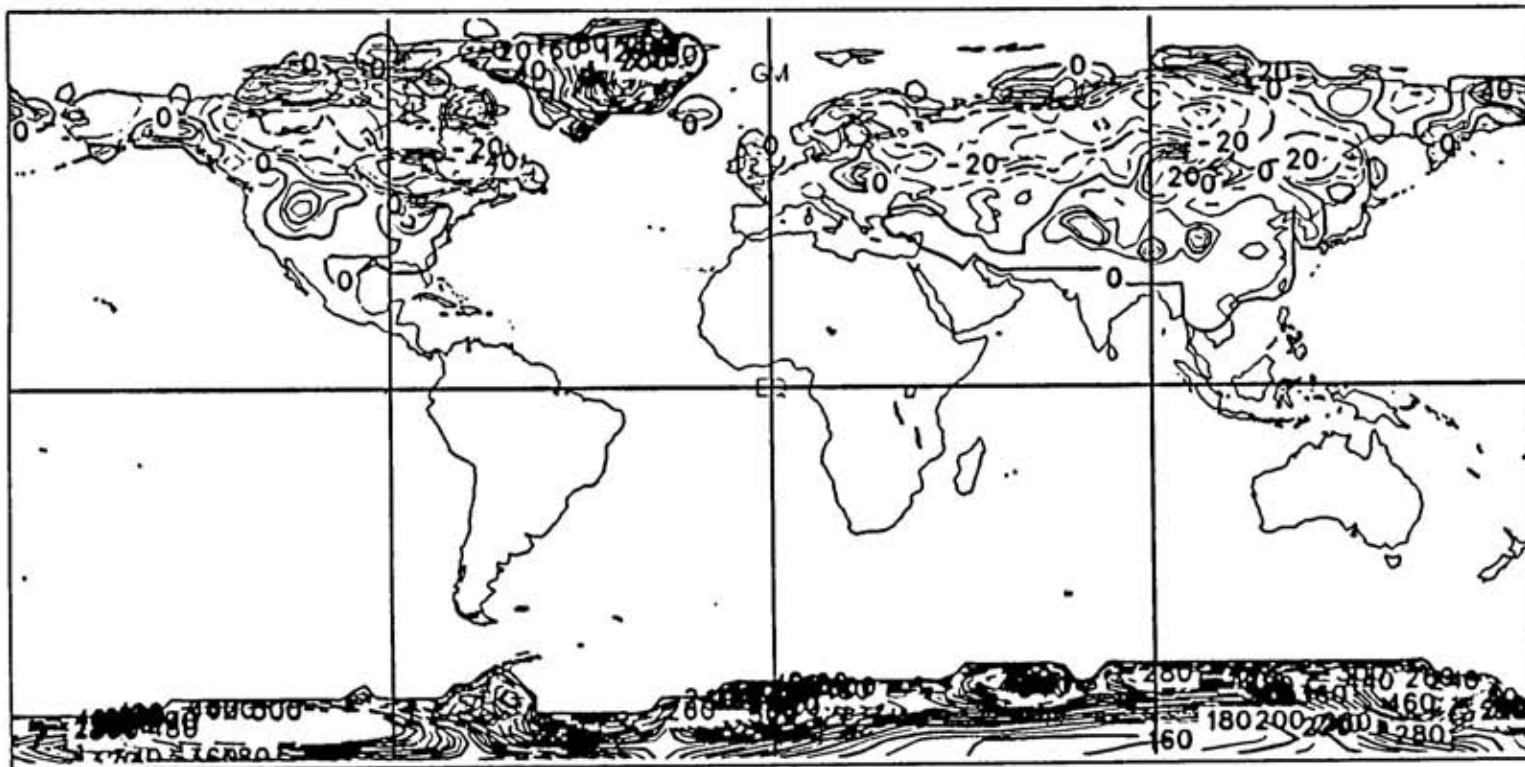
Vegetation influences the snow mass budget through interception of precipitation and shielding of the snow cover from the atmosphere and incoming radiation fluxes. In most land surface schemes, the treatment of vegetation is very simple. Some consider plant canopies as affecting only the albedo of the surface; they may or may not account for the effects of partial burying by snow. The most sophisticated models treat the canopy as thermally separate from the ground, explicitly solving for the energy fluxes between the vegetation and the underlying surface. Almost all, however, approximate the canopy as being comprised only of the single most frequently occurring vegetation type; all assume that the canopy is spatially homogeneous.

Two year-long runs were carried out using the Canadian Climate Centre GCM coupled in turn to different land surface schemes. One, a second-generation model called CLASS (Canadian Land Surface Scheme) treats snow cover explicitly alongside a three-layer soil model and a thermally-separate canopy; the other uses the force-restore method for soil temperatures and the bucket model for soil moisture. The resultant snow depth anomalies for the two runs (modelled-observed) are shown in Figure 1. Zonally-averaged values of snow depth over land, excluding continental ice sheets, are shown in Figure 2. (The spike produced by both models at 80° N represents only a few land points at the margin of the Greenland ice sheet, where the representativeness of the few available measurements is in any case questionable.) It can be seen that although the modelled snow extents are similar, and both models underestimate snow depth, the errors associated with the old scheme are much larger. This is directly attributable to its treatment of the ground thermal regime. Firstly, because the temperature of the surface is not allowed to go below zero until all of the soil moisture is frozen, any snow falling on the surface before this has occurred is melted and either added to the soil moisture or allowed to run off. Since the soil profile may in this model be up to two metres deep, this introduces a considerable lag into the appearance of the snow pack. In contrast, CLASS allows the surface to freeze and snow to accumulate before even the first soil layer is entirely frozen. Secondly, in the old scheme, the formulation used does not allow frozen soil moisture to thaw until almost all of the snow is gone. Energy which would normally be used to thaw the ground is thus used instead to effect an anomalously fast disappearance of the snow pack. (This simple treatment of snow cover and frozen water is common to most GCM land surface schemes, including those which were used in the production of the doubled-CO<sub>2</sub> scenarios being used in a variety of climate change impact studies.)

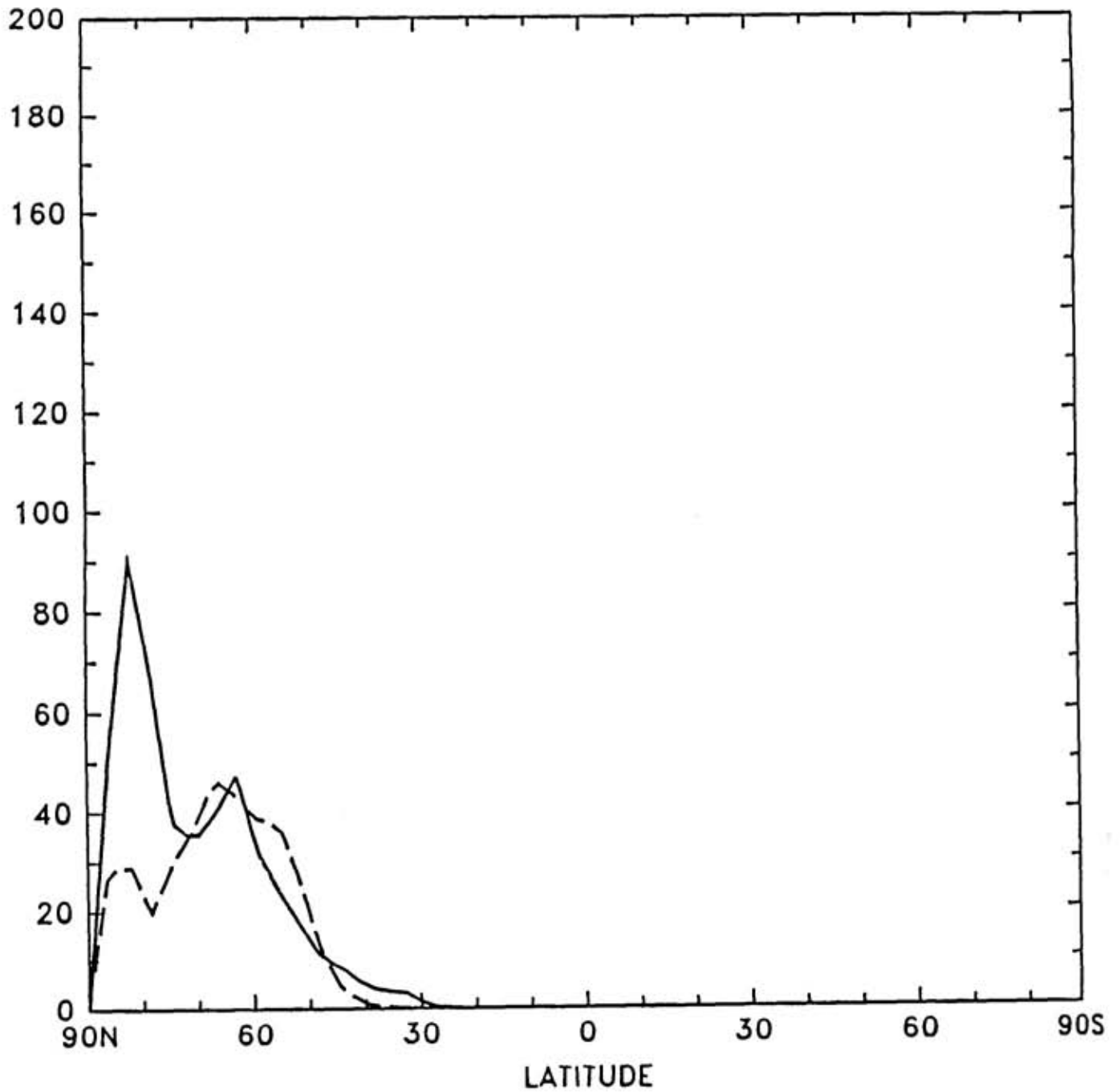
The fact that the old scheme does not allow surface temperatures to rise above/fall below freezing until all of the soil moisture has thawed/frozen also leads to anomalously large energy exchanges at the surface. While the soil is freezing and the surface temperature is artificially held at 0° C, highly unstable conditions are set up above the surface and large upward turbulent energy fluxes take place. When all of the soil water is frozen, there is nothing to prevent surface temperatures from falling to anomalously low levels, since the restraining effect of unfrozen subsurface layers is absent. When the soil is thawing and the surface temperature is again held at 0° C, stable conditions are set up and no turbulent transfers of heat away from the surface take place until the frozen soil moisture has melted completely, again resulting in large ground heat fluxes, this time into the surface. Clearly, this means that permafrost cannot be modelled. It also means that the



**Figure 1a.** Snow anomaly (modelled - observed values) in cm for the season of December-January-February -- CLASS.

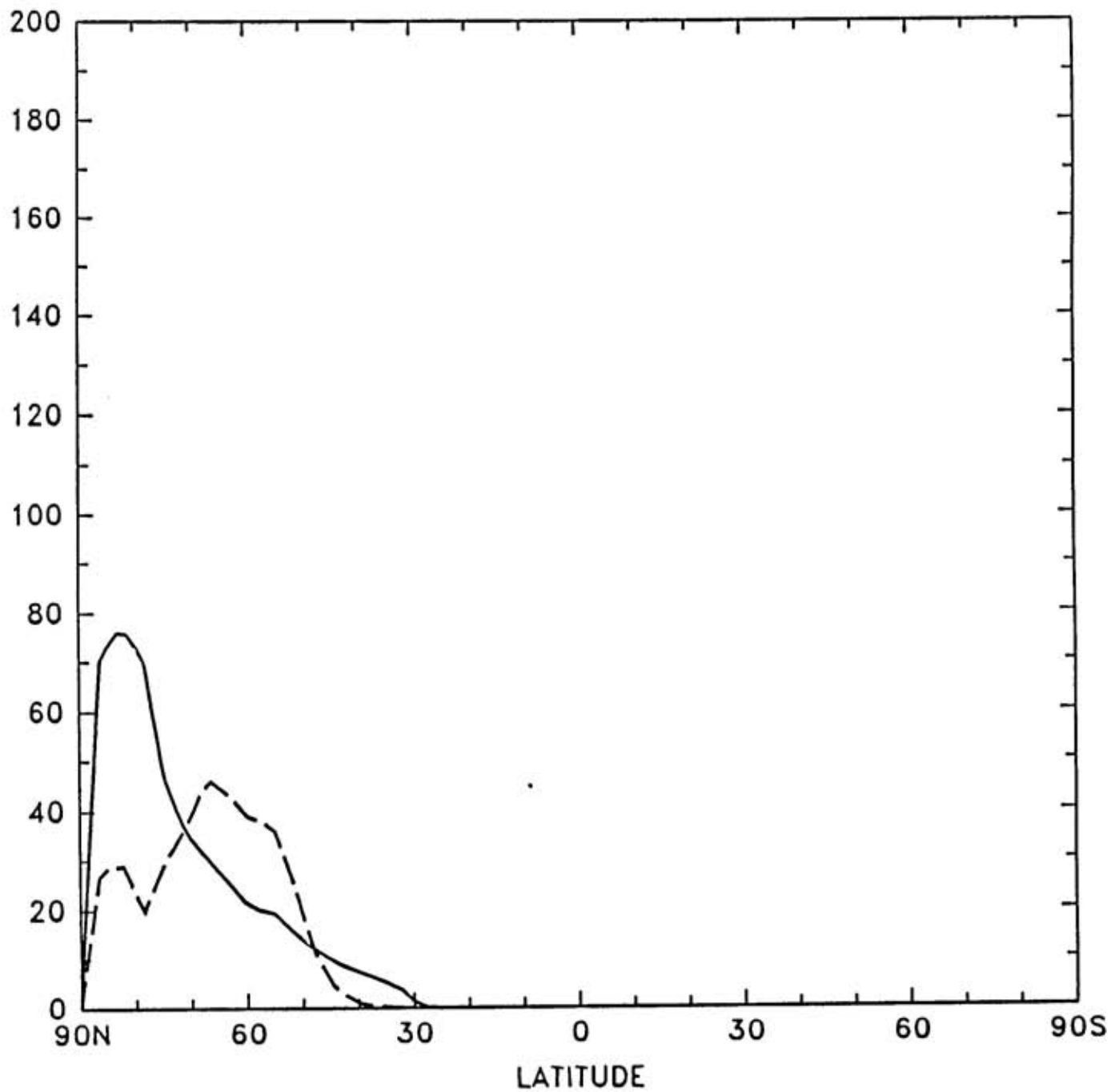


**Figure 1b.** Snow anomaly (modelled - observed values) in cm for the season of December-January-February -- old scheme.



**Figure 2a.** Zonally averaged snow depth produced by CLASS (over land areas only, excluding continental ice sheets) in cm. Solid lines indicate modelled values, dashed lines observed values.





**Figure 2b.** Zonally averaged snow depth produced by the old scheme (over land areas only, excluding continental ice sheets) in cm. Solid lines indicate modelled values, dashed lines observed values.

summertime surface latent heat flux is very likely to be overestimated in permafrost areas, since liquid water is available for evaporation throughout the soil profile.

It is evident that in order to realistically model the surface climate in areas which experience a seasonal snow cover or are underlain by permafrost, a soil model is required with several layers for which temperature, liquid moisture and frozen moisture are prognostic variables, and snow cover needs to be modelled as a layer distinct from the underlying soil. However, even in models which incorporate such features, further information will be required before snow can be reliably modelled. Process studies are needed to define curves of fractional snow cover versus snow mass for different terrain types, and to quantify the snow masking and snow interception characteristics of various vegetation types. Finally, global climatologies of monthly snow mass and areal extent must be developed, so that the performance of snow modelling schemes can be tested.

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# Glacial Start and Global Warming: What to Watch

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## Abstract

Contrary to the expectations of most predictive models of the CO<sub>2</sub> impact, the northern North Atlantic and the surrounding lands cooled between 1945-1986 and the temperature gradient between the low and high latitudes increased. The geographic and seasonal distribution of recent temperature anomalies shows a close resemblance with the 26-30 years NCAR model run of Washington and Meehl (1989) in which the climate responds to the gradual CO<sub>2</sub> increase by 1% annually. The relative cooling of high northern latitudes opposed by warming of low latitudes and the relative cooling of the middle northern latitudes opposed by a relative warming in spring, show qualitative parallels with current insolation trends at the top of the atmosphere. It is speculated that the current warming of low latitude oceans accompanied by the cooling of northern North Atlantic may represent a pattern favoring increased snow accumulation and glacier growth in northeastern Canada and Scandinavia in which case the continuing global warming possibly assisted by the CO<sub>2</sub> rise would not be necessarily incompatible with a glacial onset.

It is thus of utmost importance to monitor separately and in detail the cryospheric as well as other weather related and oceanic variables in the Arctic and in the sectors surrounding the northern North Atlantic. The developments in that zone will show whether the cryosphere will eventually retreat, giving way to the model predicted superinterglacial, or whether it will grow, marking the onset of the CO<sub>2</sub> assisted glacial.

## Introduction

No other element of climate system plays so many different roles as the water. As a liquid it effectively stores and redistributes solar energy throughout the world. As a vapor it acts as a potent greenhouse gas. As a solid, whether in the atmosphere or on the ground, it reflects much of the incoming solar radiation effectively cooling the surface. It keeps the atmosphere isolated from the underlying water bodies and therefore dry and cool. The transformation of water from its heating into a cooling form makes the ice a key player in climate change.

What will be the role of snow and ice in the climate shift expected to result from the increasing man-made greenhouse gas concentrations in the next century? What role will it play in the transition from the current interglacial into a future glacial climate? Where are the important areas to watch and the important seasons to monitor?

Limited sophistication of climate models currently available does not yet provide reliable answers. When it comes to water in whatever form, the model results differ widely. Model predictions of increasing CO<sub>2</sub> impact disagree with the geographic and seasonal pattern of the currently observed trends. While the low latitudes are warming, there are no clear signs of a uniform retreat of snow and ice in the high northern latitudes which according to most models should herald the advent of the CO<sub>2</sub> superinterglacial.

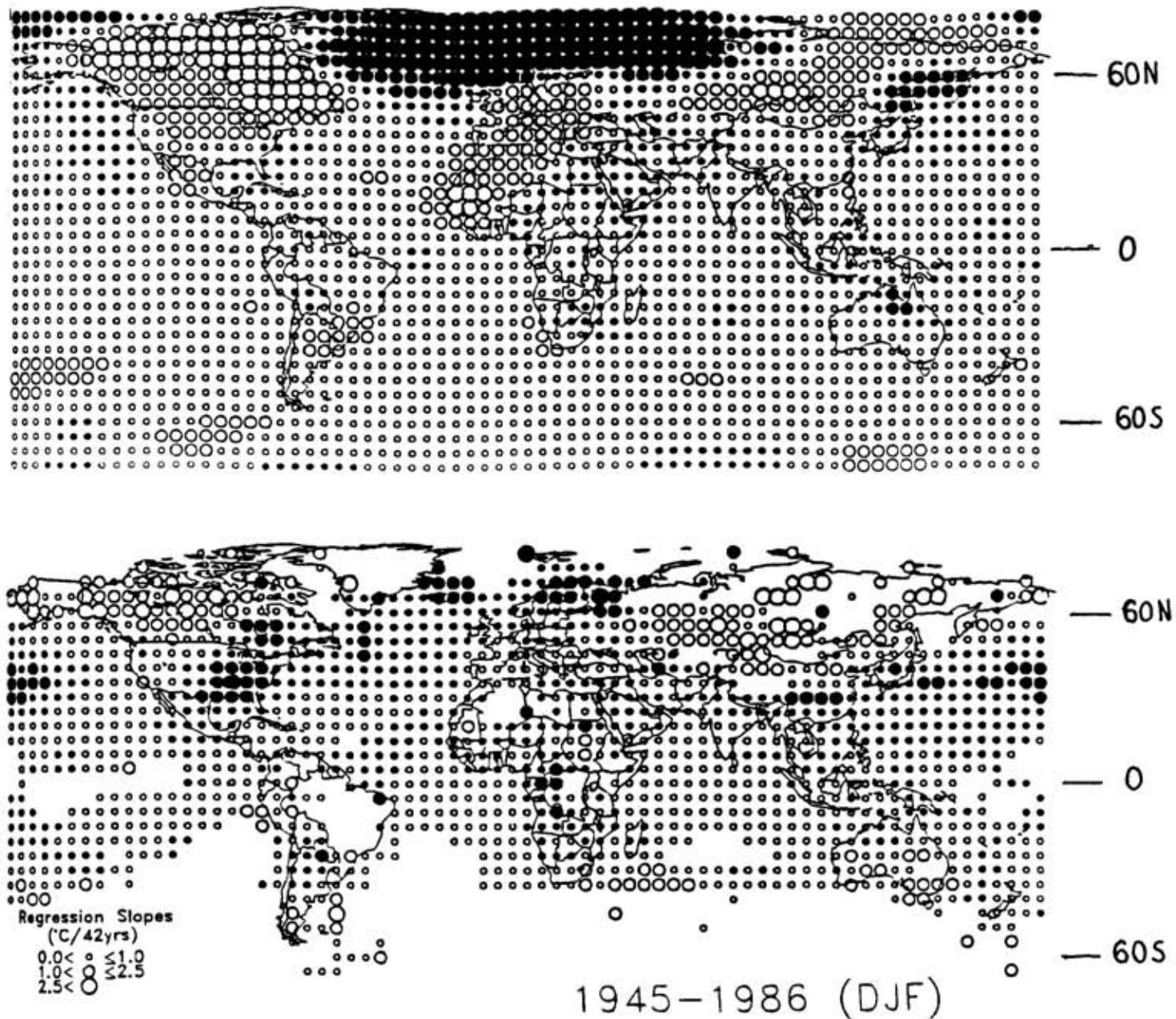
## Geography of Current Surface Temperature Trends

The surface air temperature in the post-war era, the time of the most rapid and most substantial CO<sub>2</sub> increase, and the time of rising global mean temperatures show areas of large negative anomalies. Linear trends computed from the data of Jones et al. for the interval 1945-1986 (Kukla et al., 1992), show for winter (D,J,F) a cooling of up to 1.5 to 2°C over eastern North America, the northern North Atlantic, northern Europe, and the western and central Pacific (Fig. 1). Strong warming was observed over Alaska, northwestern Canada and north central Asia, with a maximum in northern Siberia. The oceans in low northern latitudes and in the Southern Hemisphere show mostly warming of up to about 1-1.5°C. The spring trends, although less expressed, resemble the winter pattern, with pronounced warming in northwestern North America and northern Asia but cooling in the North Atlantic and northwestern Pacific. In summer (J,J,A) the Atlantic and the northwestern Pacific cooled and so did also Europe and the Mediterranean Basin (Fig. 2). Cooling was most extensive in autumn (S,O,N) and covered much of the North Pacific, North Atlantic and North America. Warming prevailed over the southern oceans and lands, north central Asia and low latitude northern oceans. The warm anomalies in fall are much weaker and cover smaller areas than in other seasons. Large interannual variability affects the record at every grid point and in all seasons.

Comparable temperature changes were reported earlier (Jones et al., 1986; Oort et al., 1987; Walsh and Chapman, 1990; Karoly, 1990; Newell and Hsiung, 1990; Bunker, 1980) and the cooling of the northern North Atlantic and northeast Canada continued through the early 1990's (World Meteorological Organization, 1992). Principal aspects of surface anomalies are reflected in tropospheric thickness between 1000 and 300 mb during the 1977-1986 interval (Weber, 1990). Negative anomalies are located over the North Pacific, North Atlantic and the Canadian and European sectors of the Arctic and prominent positive anomalies over the tropical and subtropical Pacific. The temperature difference of the atmospheric column between subtropics and the middle latitudes increased by more than a degree and the zonal mean upper air temperatures between 850 and 300 mb levels (Angell, 1990) show large increase in the low latitudes and only a small change over the Arctic. As a result, the meridional temperature difference of the atmospheric column increased significantly.

Changes of precipitation and salinity were also observed. The mean zonal annual precipitation in the middle and high northern latitudes shows an oscillatory increase since at least the 1920's (Diaz et al., 1989). Precipitation increased notably in northeastern Canada during the sixties, seventies and eighties (Danard et al., 1990) and also in the high latitudes of Europe and Asia (Bradley et al., 1987). The extent of sea ice in the Hudson Bay increased between the start of satellite monitoring in 1973 and the late 1980's (Parkinson, 1989). Snow accumulation in the southern part of the Greenland ice was higher in recent years (Zwally, 1989) and the mountain glaciers in northern Norway are expanding (Fairbridge and Möller, pers. commun.). No change in the mean sea ice extent was found in the East Greenland, Labrador and Barents Seas, areas of strong negative sea surface temperature and salinity anomalies (Walsh and Chapman, 1990).

Surface waters near Iceland and in the Labrador Sea freshened considerably between 1968 and 1972 during the so-called "Great Salinity Anomaly" which was accompanied by exceptionally extensive sea ice and a temporary interruption of deep water convection (Mysak and Manak, 1989). Formation of deep water in the East Greenland Sea slowed down further during the 1980's as inferred from radioactive tracers, and the production of dense saline water was reduced by 80% (Schlosser et al., 1991). The strength of northerly winds and storminess in the east Atlantic between 35 and 65°N and 20°W to 10°E have



**Figure 1:** Trends of winter (DJF) mean surface air temperature during the 1945-1986 interval (bottom panel) compared with the departures of the 26-30 year NCAR transient model output from the control run (upper panel). Full circles show cooling, open circles warming. The small circle is for departures equal or less than 1°C, the large circle for over 2.5°C. The concentrations of CO<sub>2</sub> in the model increased by 1% annually for 30 years, approximately comparable to the increase of combined greenhouse gases in the 1945-1986 interval.

Observed trends were computed from the grid data of Jones et al. (1989). Model results are those of Washington and Meehl (1989). Recomputation of both data sets on comparable grid courtesy of G. Meehl, R. Knight and T. Karl.



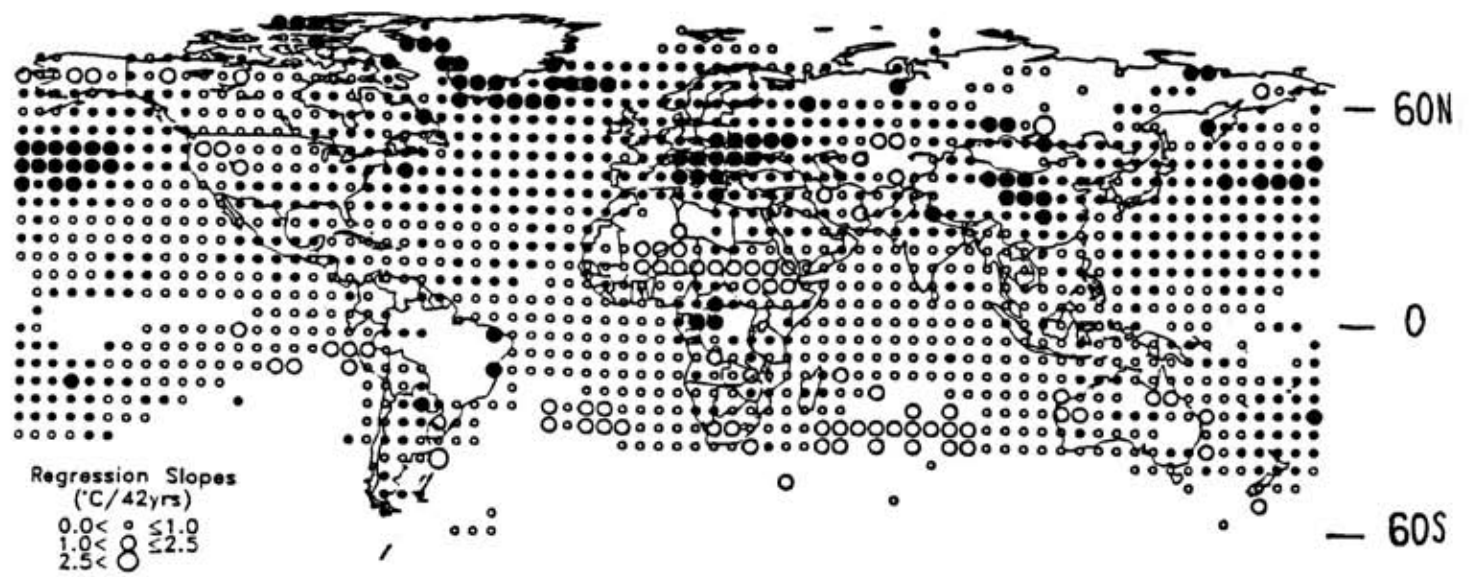
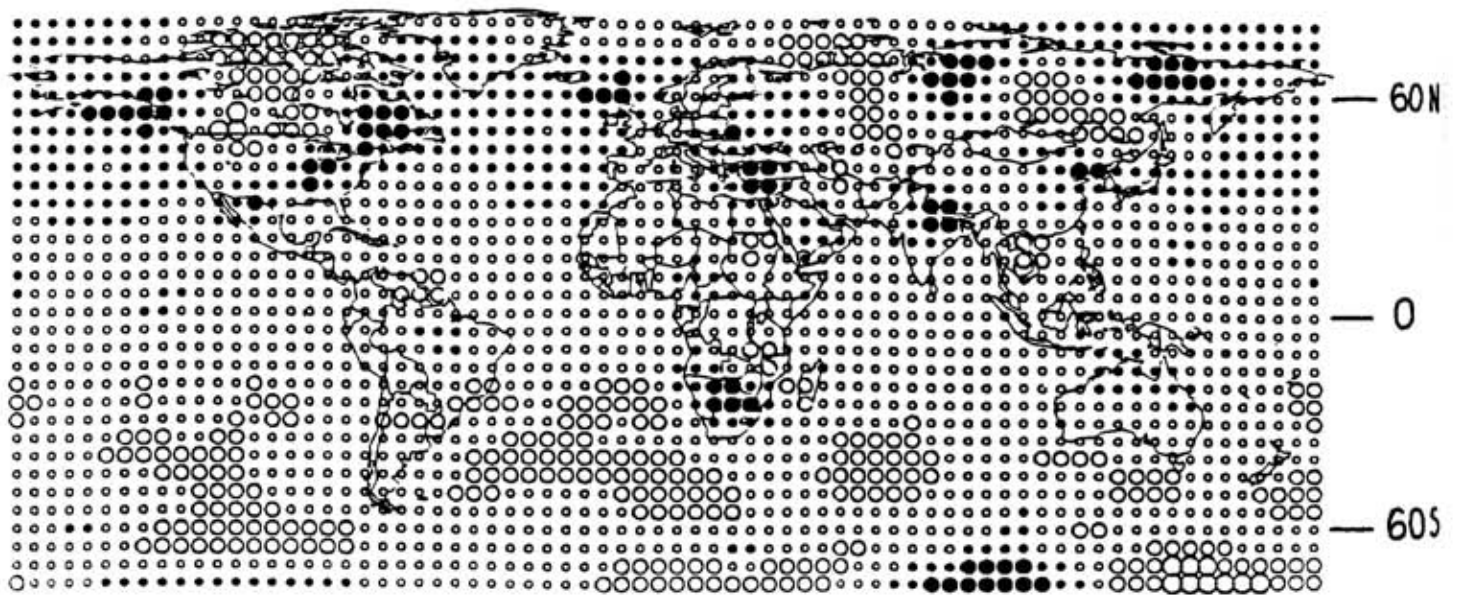


Figure 2: Same as Figure 1, but for summer.

increased between 1946 and 1980 (Dickson et al., 1988). As pointed out by Aagaard and Carmack (1989), salinity levels are now close to a critical threshold at which the sinking motion of water in the area and the import of warm waters into the Norwegian and Barents Seas may be greatly reduced.

To find out whether the mean zonal temperatures of the high latitudes increased in the 1945-1986 interval as expected from the CO<sub>2</sub> impact, Kukla et al. (1992) reduced the anomalies to areally weighted zonal means polewards of 65°, between 45°-65°, 25°-45°, 5°-25°, and between 5°N and 5°S. The resulting seasonal linear temperature trends for land and ocean for the 1945-1986 interval were expressed in °C per century for the Northern Hemisphere winter (DJF), spring, summer and autumn. Figure 3 shows that:

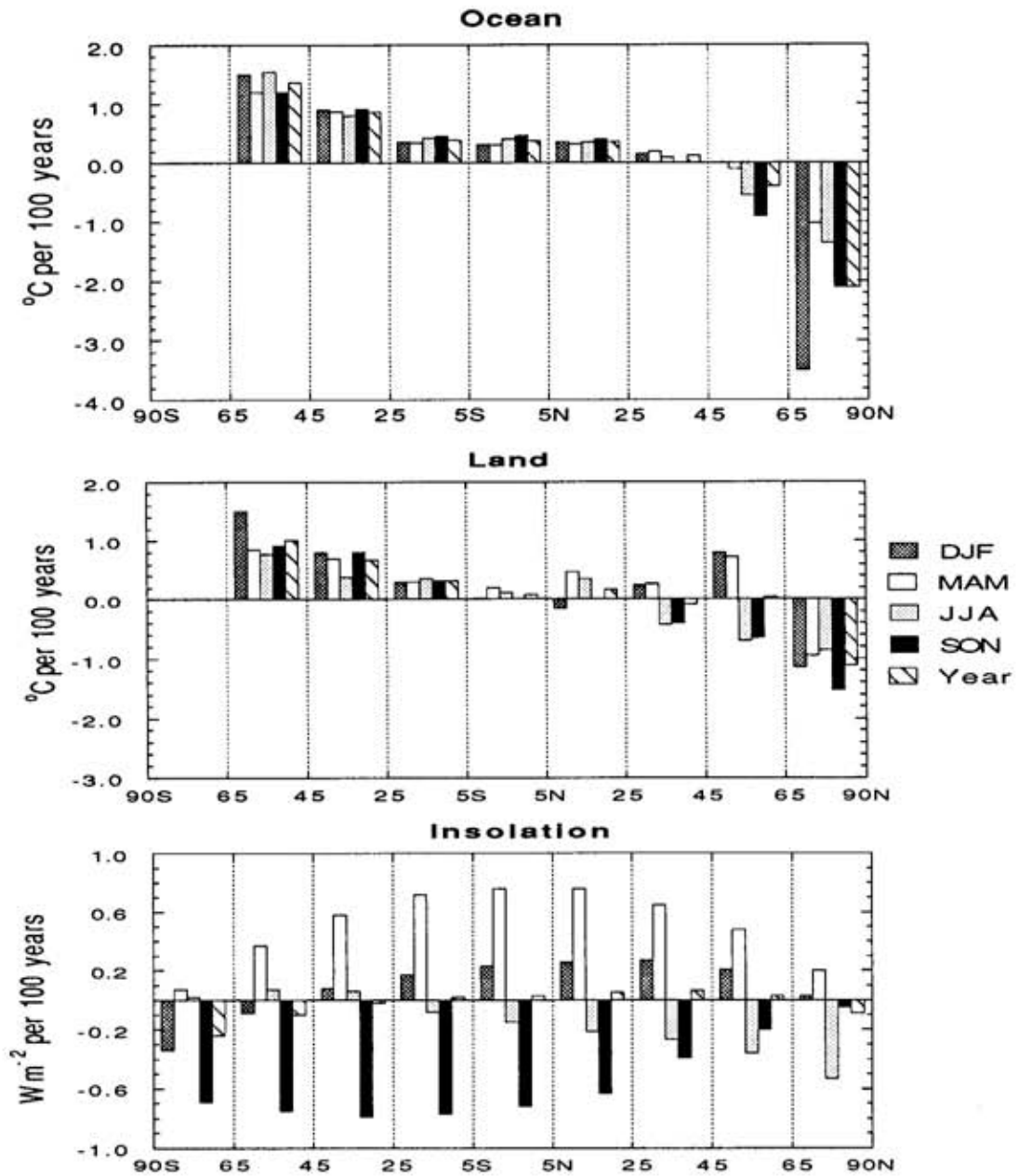
- 1) the Arctic (65-90°N) cooled in all seasons over both the land and the ocean. The cooling was strongest in winter over the Atlantic Ocean,
- 2) the northern low latitudes and those parts of the Southern Hemisphere, where data were available, warmed in all seasons, dominating the global trend,
- 3) the surface temperature gradients between the high and low northern latitudes and between the Northern and the Southern Hemisphere increased in all seasons,
- 4) the North Atlantic and the northwestern Pacific cooled more than the surrounding land, and as a result the summer temperature gradient between the land and the oceans increased in the middle and high northern latitudes,
- 5) the temperature trends north of 25°N show large seasonal differences with strongest cooling in autumn and summer and warming or lesser cooling in winter and spring,
- 6) warming trends over the oceans south of 45°N showed only small seasonal differences.

### **Comparison With Models**

It is obvious that the observed features are not what one would expect from a linear interpolation of the 2 x CO<sub>2</sub> minus 1 x CO<sub>2</sub> equilibrium climate models. Such interpolations call for a preferential warming of the high latitudes (especially during winter) with a relatively modest temperature increase in the low latitudes. As a result the meridional temperature difference between the low and high latitudes should be decreasing. Instead, the low latitudes in the 1945-1986 interval warmed more uniformly than the high latitudes and the meridional gradient between the equatorial zone and the Arctic increased.

While the recent temperature anomalies do not show much in common with the projections based on equilibrium models, they do show features in common with the output of three recent studies, which are:

- 1) the NCAR model prediction of the climate response to the gradual CO<sub>2</sub> increase by 1% for 30 years (Washington and Meehl, 1989),
- 2) trends of the current insolation shifts on top of the atmosphere (Kukla et al., 1992),
- 3) conceptual model of glacial onset of Kukla and Gavin (1992),



**Figure 3:** Zonal mean seasonal surface air temperature trends in the 1945-1986 interval expressed as  $^{\circ}C/century$ , areally weighted for the belts bounded by latitudes  $65^{\circ}$ ,  $45^{\circ}$ ,  $25^{\circ}$  and  $5^{\circ}$ , separately for the oceans and land. Compared with corresponding current mean seasonal insolation trends on top of the atmosphere in  $Wm^{-2}$  per century. Data from Kukla et al. (1992), Jones et al. (1989) and Berger and Loutre (1991).

## The NCAR Transient Response Model Prediction

The NCAR study is based on the coupled ocean-atmosphere model which includes ocean circulation. In the model, the transient response of climate system to the gradual increase of CO<sub>2</sub> by 1% annually has been analyzed (Washington and Meehl 1989). Such representation of the CO<sub>2</sub> rise is more realistic than the equilibrium adjustment to instant doubling. It approximately corresponds to the actual rate of increase of combined greenhouse gases during the post-war interval. The model output should be thus roughly comparable with the temperature change observed between 1945 and 1986. In order to assess the greenhouse impact, the model results have been compared to the control run in which the CO<sub>2</sub> atmospheric concentration has been kept constant at 330 ppm. With the transient response, the five year average ending at year 30, by which the CO<sub>2</sub> rose by approximately 30%, showed a 0.7°C increase of the global mean temperature. However, large cooling areas appeared in the northern North Atlantic, the surrounding lands and in the northwest Pacific. Pronounced warming is observed in the model in the continental interiors in Alaska, Canada and central Asia. The temperature changes are strongest in northern winter (Fig. 1) and weakest in summer (Fig. 2). The cooling pattern is associated in the 26-30 year model run with a weakening of the Icelandic Low and its shifting toward England, intensified northerly winds in the western part of the northern North Atlantic, with increased precipitation and with freshening and cooling of the upper ocean. This in turn slows down the modeled deep water convection. Similar intensification of northerly winds occurs in the North Pacific where it is associated with a shift of the Aleutian Low.

The distribution of temperature anomalies in years 26-30 of the transient model has several features in common with the observed trends. In both, the model and the observations, the two areas of strongest winter warming are found over northwestern North America and central and northeastern Asia while the North Atlantic, the Barents and Kara Seas and the northwest Pacific cool. In summer much of the North Pacific cooled both in the model and in the real world. The belt of the most pronounced warming is in the southern middle latitudes, and is stronger in JJA than during DJF. The warming anomaly in North America shrinks and weakens in summer compared to winter in both the model and the observations. Cooling of the North Atlantic is weaker in summer than in the winter, but expands over the eastern part of North America. In both the model and the observations, northerly winds in the northwestern Atlantic intensified and the salinity decreased. There are serious disagreements as well between the observations and the model. In winter, eastern North America cooled, but it warmed in the model. Northeastern Pacific and Alaska warmed but they cooled in the model. The observed winter warming in Asia is two to three times stronger as simulated by the model, and the observed warming of the low latitude oceans is also stronger than in the model. Observed cooling in the North Pacific, Northeast Atlantic and the Norwegian Sea in winter and summer is larger than in the model. The observed mean warming rates over the ocean relative to those over land are considerably higher than in the model. The principal area of the deep water convection in the model is at lower latitude than as observed in the real world.

Although the seasonal and geographic pattern of the postwar surface air temperature trends differs from the NCAR model simulation in years 26-30 in many respects, it has more features in common with this trial than with any other CO<sub>2</sub> impact model published to date. It is encouraging to see such parallels between one sampling period of the transient model and the observed post-war trends, but given the large variability of observed surface temperature anomalies in time and space and the large variability of NCAR model as well, it can not be discounted that the match is entirely fortuitous.



## The Insolation Change

Past large scale climate changes have been shown to be highly correlated with periodicities of the earth's orbit (Hays et al., 1976; Kukla et al., 1981; Imbrie et al., 1984). On that basis it is widely accepted that on the geological time scale a gross cooling trend is imminent (Kukla et al., 1972; Imbrie and Imbrie, 1979). While the past climate shifts broadly followed the slow insolation lead, they apparently occurred in several steps of relatively rapid change separated by intervals of relative stability (Beaulieu and Reille, 1989; Mix and Ruddiman, 1985).

Recently, Kukla et al. (1992) compared the ongoing changes of insolation on top of the atmosphere due to orbital perturbations with currently observed temperature trends to determine whether the current climate developments may be related to the mechanism of orbitally controlled interglacial-glacial transition.

The authors computed seasonal trends of zonal mean insolation at the top of the atmosphere (TOA) from the differences between the insolation fields 1,000 years ago and 1,000 years in the future, as given by Berger and Loutre (1991). The results were expressed as a difference of mean daily insolation income at the top of the atmosphere in  $Wm^{-2}/100$  years and areally weighted for the same zones as the observed zonal surface temperature means mentioned earlier (Fig. 3). The total annual insolation to the top of the atmosphere decreased north of  $65^{\circ}N$ , but increased in the low latitudes. Over the globe it remained practically unchanged. The precession induced seasonal trends are much more pronounced. The mean seasonal insolation is currently decreasing at every latitude during the northern autumn (SON) and increasing everywhere in spring (MAM). A decrease is also found north of  $20^{\circ}S$  in summer and an increase north of  $50^{\circ}S$  in winter.

The insolation shifts have three qualitative features in common with the observed temperature trends, namely, (1) an annual decrease in high northern latitudes and increase in low latitudes, (2) increase of the meridional gradient between the low and high northern latitudes, and (3) relative decrease in autumn and increase in spring over northern lands and oceans (Fig. 3).

In quantitative terms, the changes of insolation are small and almost negligible when compared to the impact on surface radiation budget of the day-to-day fluctuations of cloud cover, atmospheric humidity, precipitation, and so on. The peak TOA insolation increase over the 1945-1986 interval is only  $0.18 W m^{-2}$  in April at  $10^{\circ}N$  and the peak decrease is  $0.25 W m^{-2}$  at the South Pole in November.

Unquestionably, the current temperature trends were not directly caused by orbital forcing. The bulk of the observed short term fluctuations should be instead viewed as an internal oscillation along a stationary long term mean (Lamb, 1972 and van Loon and Rogers, 1978). The partial qualitative agreement between the ongoing temperature and insolation trends thus was, with high probability, only coincidental. The authors however ask whether such coincidences, if repeated, could influence the climate on the time scale of centuries and millenia. They propose that on the longer run, the regional climate developments could be systematically amplified when in phase with the ongoing insolation changes and suppressed when out of phase.



## The Conceptual Model of Glacial Onset

The conceptual model of Kukla and Gavin (1992), is based on the analysis of insolation trends occurring at times of the past gross transitions of global climate from a warm to a cold stage and vice versa. The warm to cold transitions identified in the radiometrically dated paleoclimatic evidence were centered at about 117, 97 and 73 ka. They were marked by peak spring increases of insolation on top of the atmosphere to the low latitudes, by the peak decrease of insolation to high northern latitudes in northern summer (July, August, September), and by a decrease to the high southern latitudes in southern spring (October, November). Opposite pattern was found for cold to warm transitions.

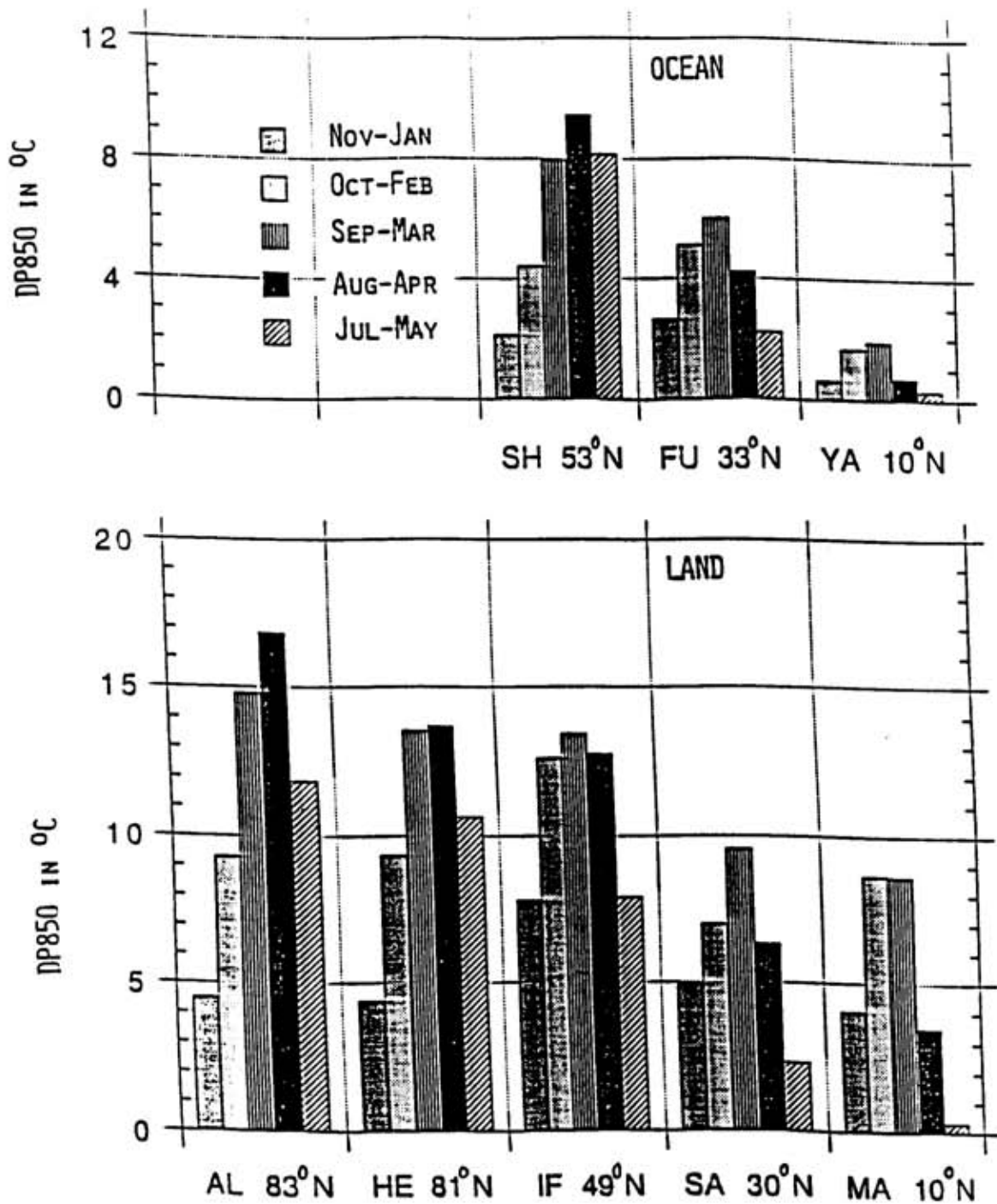
Based on the observations of the current seasonal cycle Kukla and Karl (1992) and Kukla and Gavin (1992) identified two mechanisms which can respond to the Milankovitch's precessional signal. Both are related to an approximately two month lag of the hydrosphere behind insolation forcing.

They are (1) the seasonal cycle of greenhouse forcing of surface temperature, controlled by the variation of atmospheric water vapor concentrations and (2) the gradual heat storage in surface ocean during spring and summer followed by heat release in autumn and winter.

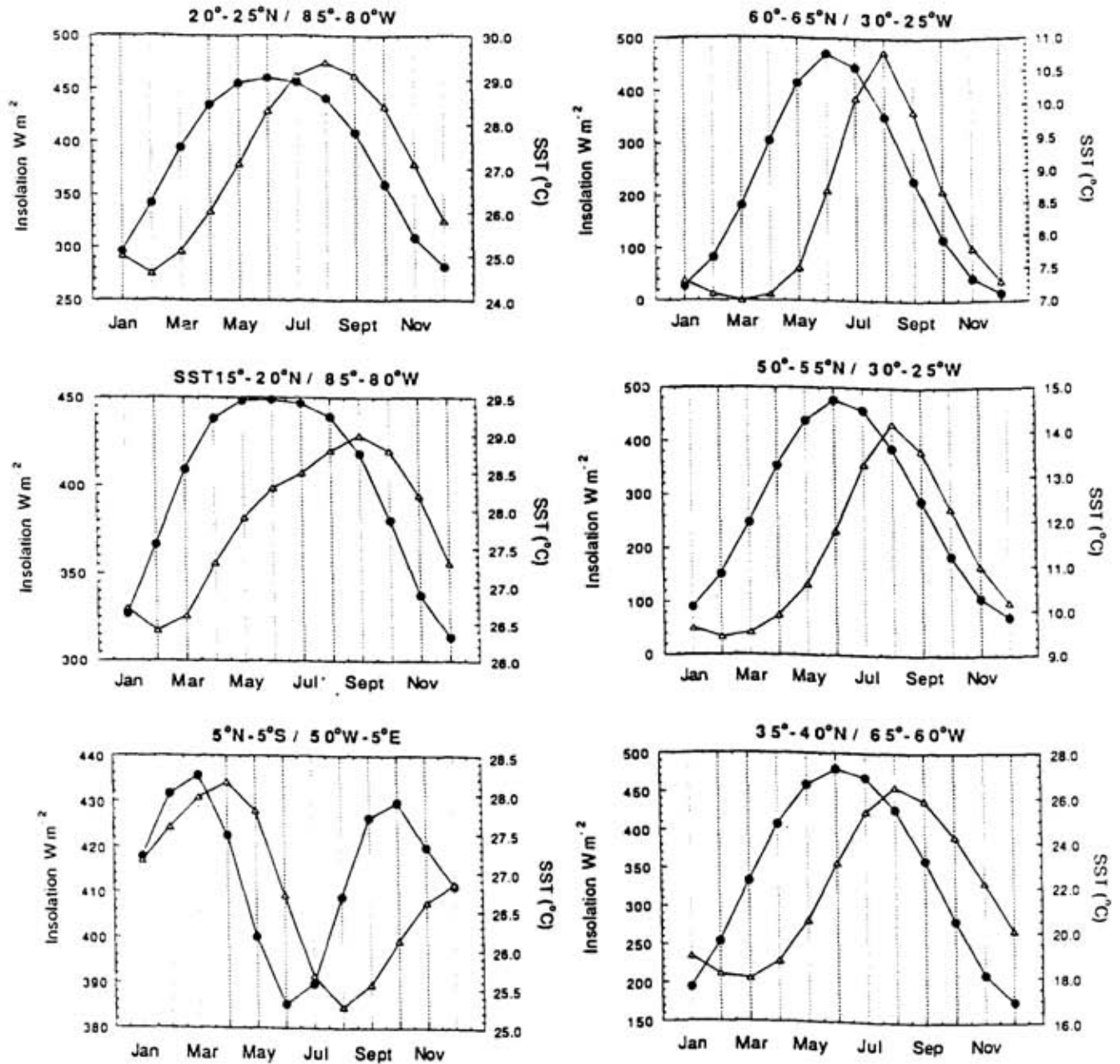
Water vapor (precipitable water) concentrations in the high and middle latitudes and especially over land are considerably lower in spring, than under the same sun configuration in autumn. Thus the heating efficiency of fall insolation arriving at time of stronger greenhouse effect is greater than in spring (Fig. 4). Fall insolation increase at the expense of spring decrease should lead to surface warming, the opposite to surface cooling (Kukla and Karl, 1992).

The gradual heating of the Northern Hemisphere oceans between February and September follows the insolation lead with a 2-3 month lag (Fig. 5) followed by a dynamically induced cooling in October through January. Increased insolation in spring is therefore expected to lead to warmer waters available at the time of autumn surface overturn and an increased water vapor transfer to the atmosphere. In their conceptual model of the interglacial-glacial transition (Kukla and Gavin, 1992) propose that (cf. Fig. 6):

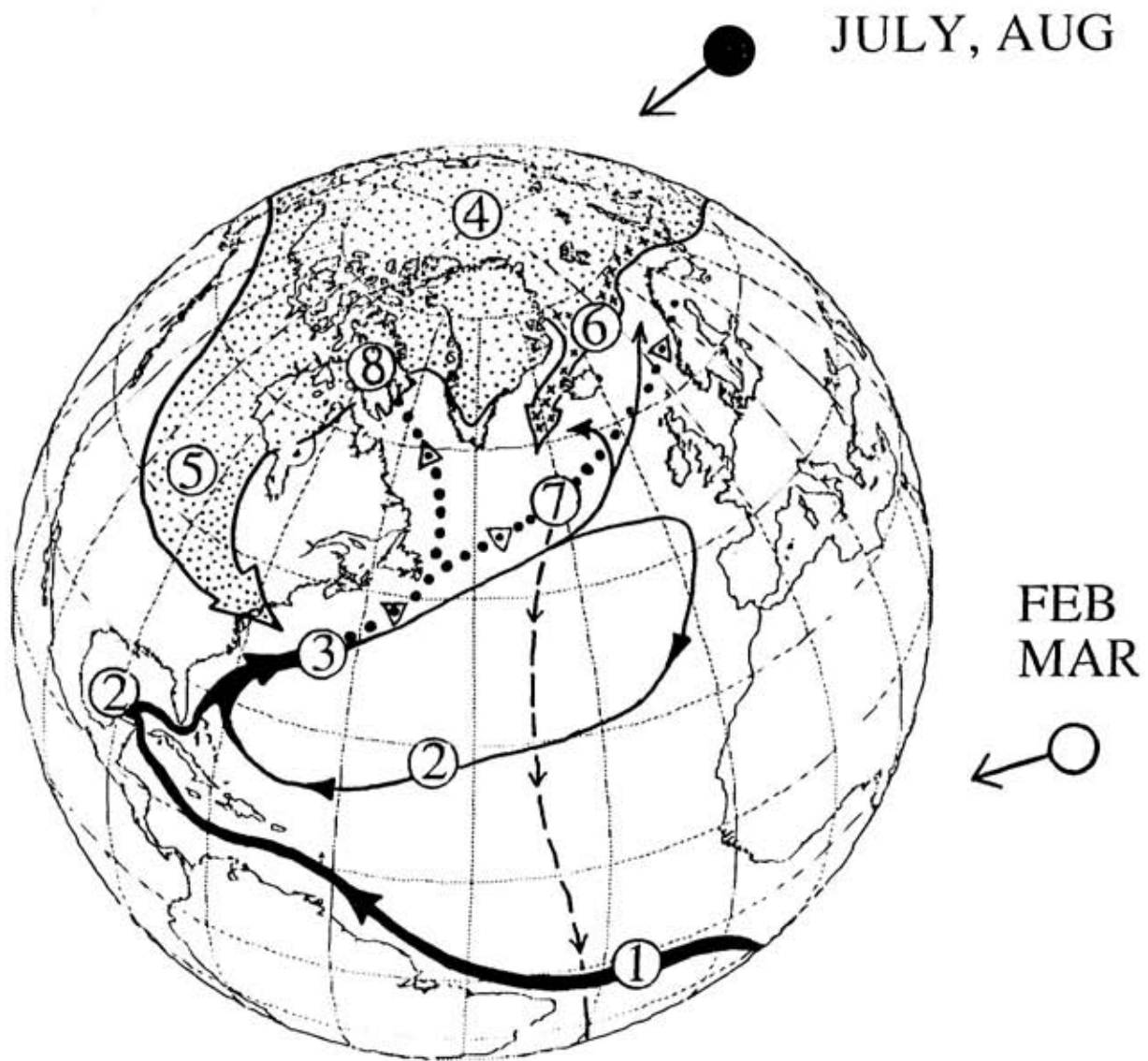
- 1) More intense insolation in spring warms the Northern Hemisphere oceans including the tropical and subtropical Atlantic. Currents carry the warmer water into the head zone of the Gulf Stream off North America. (It takes approximately half a year for subtropical surface waters to reach this area.)
- 2) Decreased July and August insolation leads to a reduction and eventual cessation of summer snow melt in the Arctic. Year-round dominant high pressure regime and frequent invasions of polar air masses into the middle latitudes follow. Decreased fall insolation to the mid-latitudes (at the time of maximum seasonal greenhouse effect over land, in the sense of Kukla and Karl, 1992), leads to the expansion of snow covers and promotes deeper southward penetration of cold-dry continental air outbreaks. The frequency of the so-called "Greenland Above" circulation type, which Crowley (1984) suspected to favor glaciation increases.



**Figure 4:** Difference of the dew point (°C) at the 850 mb level taken as a proxy of precipitable water between the months paired by approximately equal TOA insolation at stations: ALert, HEjsa, Int. Falls, San Antonio, MALakal, SHeyma FUnchal and YAp. Station latitude also indicated. Centered at 21st of each month. From Kukla and Karl (1992). Differences in precipitable water (and related greenhouse forcing), for the same TOA insolation input are larger over land than over the ocean and peak in late summer (August and September as opposed to April and March).



**Figure 5:** Insolation on top of the atmosphere (full circles) and the mean sea surface temperature (SST, open triangles) in the Atlantic Ocean during the 1951-1980 interval (data from Bottomley et al., 1990 and Berger and Loutre, 1991). SST lags about 2-3 months behind TOA insolation. Ocean cooling in the second half of the year north of 15N lags only 1 to 1.5 months behind the insolation wave, being strongly affected by dynamic mixing of the upper ocean.



**Figure 6:** Conceptual model of the interglacial-glacial transition in the North Atlantic area. Increased insolation in spring warms ocean currents in the equatorial Atlantic (1) and the subtropical belt (2). Warmer water reaches the zone of intense fall and winter storminess and evaporation (3). Decreased summer insolation in the Arctic (4) leads to the year-round maintenance of high albedo and dominance of anticyclonic circulation. More frequent and intense polar air outbreaks (5) interact with the ocean off N.E. America (3), increasing the storminess and the fall and winter evaporation. This in turn leads to increased precipitation and snow accumulation in N.E. Canada (8) and decreased salinity in northern North Atlantic. Increased freshwater runoff to the Arctic results in increased sea ice export from the Arctic (6) into the Conveyor Belt headwaters (7) and eventual shutdown of the thermohaline circulation (Kukla and Gavin, 1992).

- 3) Increased water vapor gradient between the drier continental and Arctic air and warmer ocean in the area off northeast North America causes intensified storminess (especially in autumn and winter), increased evaporation and energy loss from the ocean, as well as increased precipitation inland and over the deep water production zones of the northern North Atlantic. (The zone between 35 and 40°N and 65 to 70°W indicated in Figure 6, shows today the highest ocean-air energy transfers throughout the year.)
- 4) A precipitation increase in northwestern Canada in autumn and winter supports the build-up of glaciers and an increase of runoff into the Arctic Ocean, strengthening the salinity stratification of the upper Arctic waters. This results in an increased production (Untersteiner, 1961) and export of sea ice into the East Greenland Sea where it affects the deep water production (Mysak and Manak, 1989). In combination with increased precipitation over the ocean, a salinity of the upper ocean decreases and the zones of thermohaline downwelling relocate south or cease to operate. In either case, the conveyor belt in its present form ceases to exist.

The Southern Hemisphere in the model is considered less sensitive to insolation changes than the northern one because of small portion of land, dominant zonal as opposed to meridional oceanic circulation, and small seasonal SST range. However, the insolation decrease in October and November in high latitudes is expected to delay the dissipation of subantarctic sea ice and, in phase with the changes in the Northern Hemisphere, reinforce the cooling. Also, the shift of the deep water circulation can transmit to the Southern Hemisphere the northern cooling signal.

In this conceptual model, not yet tested by general circulation models, the stage of the warm to cold transition is set by the gradual insolation decrease in summer and fall (JASON, July through November season) to the northern lands, and by the northern winter and spring (JFMAM) insolation increase to the low and middle northern latitude oceans. The insolation decrease in July and August to a threshold level low enough to terminate the summer snow melt in the Arctic Basin, and the cessation of the north Atlantic deep water production in the Norwegian and possibly also in Irminger and Labrador Seas are seen as triggers capable of causing relatively rapid step-like transitions of climate from an interglacial into a glacial mode. The proposed mechanism is expected to operate through the interaction of the slowly changing Milankovitch forcing with the ongoing day-to-day weather oscillations, which, if unaffected by insolation, would otherwise fluctuate along an unchanging long term mean. The orbital shifts are seen as capable to increase the frequency and intensity of those circulation types which are in phase and suppress those, which are out of phase with the ongoing long term insolation trends.

## Discussion

In view of the increasing concentrations of the atmospheric CO<sub>2</sub>, how should the recent weather anomalies be interpreted? The current temperature, precipitation and salinity changes can be viewed as a purely stochastic oscillation of climate system and the last desperate try of nature to postpone the inevitable advent of the man-made model predicted greenhouse superinterglacial. Or the same anomalies can be viewed as an indication of longer term developments and of a possibly incomplete and deficient design of climate models, which do not represent realistically the impact of increasing gradients between land and oceans and between the high and low latitudes.

Several questions arise in this context:



- 1) Do the observed trends represent a temporary aberrations to be reversed in a longer run, or do they herald long term future developments?
- 2) In the latter case, and if supported by insolation shifts, as suggested by qualitative parallels between the zonal insolation and temperature trends, can the further drop of insolation in the Arctic intensify the cooling of the North Atlantic area, as foreseen by the Kukla and Gavin (1992) conceptual model of a glacial onset?
- 3) If supported at least in part by the CO<sub>2</sub> increase, as suggested by the parallels with the Washington and Meehl (1989) model, can the future rise of greenhouse gases intensify the cooling of the North Atlantic region? Or will the additional increase of CO<sub>2</sub> concentrations lead to the reversal of the trends and the massive retreat of the subpolar cryosphere?
- 4) If the anomalies of the 1945-1986 interval in the North Atlantic area result from a combined effect of natural forcings, man made impacts (including CO<sub>2</sub> and possibly aerosols), and the random variability of the climate system, what is the chance of making a reliable prediction of the next century climates?
- 5) Finally, the most provocative question: can the global warming, dominated by the increasing sea surface temperatures in the low and middle latitudes be compatible with the starting glaciers expansion in the high latitudes? Can a global warming be a first step toward a glacial?

None of the above questions can be today reliably answered. However, based on the available observations, paleoclimatic evidence and on our understanding of the physics of climate system, the questions are valid, and the problems raised are worthy of testing. While the needed analysis will have to encompass the whole globe, the key player of the expected changes is the cryosphere in the northern North Atlantic zone. The global mean averages of temperature and the global totals of precipitation will be of little use as indicators of the direction in which the climate system is shifting. To address the raised questions properly, the monitoring program of all climate variables, including snow and ice has to be regionalized and the modelling of cryosphere significantly improved. The following are some of the improvements needed:

***Monitoring:***

- 1) The changes of the land cryospheric variables, should be monitored separately by sectors in:
  - a) North-Eastern Canada,
  - b) Northern Norway, Sweden and Finland,
  - c) Western Greenland,
  - d) Eastern Greenland
  
- The variables to monitor should include snow cover thickness, water equivalent, duration and albedo, lake ice freezing and thawing dates, river runoff, precipitation and other conventional meteorologic surface and upper air parameters, including clouds.
- 2) The changes of sea ice extent and thickness, ice surface properties and albedo, sea surface salinity and temperature are needed separately by sectors in:
  - a) Central Arctic
  - b) East Greenland Sea
  - c) Barents Sea

- d) Irminger Sea
- e) Labrador Sea
- f) Hudson Bay
- g) Baffin Bay and Davis Strait

Additional meteorological variables have also to be monitored in these zones paying special attention to cloud cover.

**Modelling:**

- 1) The physical mechanism of generation, modification and dynamics of the polar air masses and their outbreaks into the middle and low latitudes should be studied in detail and reflected in a realistic way in the models.
- 2) The different impact of direct and diffuse shortwave and longwave radiation on surface environment under shallow inversions, common in the high latitudes (with snow on the ground and snow covered or snow free ice), should be studied in detail and properly represented in climate models.
- 3) Representation of the radiative impact of CO<sub>2</sub> on snow cover under shallow inversions has to be improved.
- 4) The impact of increasing air-sea temperature differences on cloud cover storminess, evaporation and eventually precipitation and runoff in the high latitudes should be analyzed in more detail and better represented in the models.
- 5) The radiative impact of CO<sub>2</sub> and industrial aerosols on the polar and subpolar cloudiness and surface climate in general requires improvement.

The above points are not to be understood as an exhaustive list of items which would bring the monitoring and analysis of the cryosphere in the study of current climate dynamics and climate change up to date. Snow cover over Alaska, Yukon territories and the American Midwest, glaciers in the Alps, lake-ice in Siberia, sea ice around Antarctica, to name just a few, are also important players in the climate system. However, if the climate has to change substantially in the near future, in one way or the other, we will see it first in the oceans, the snow, the sea ice and the glaciers of the Arctic and the northern North Atlantic and surrounding lands. Either the cryosphere of these areas shrinks, giving way to the CO<sub>2</sub> superinterglacial, or it starts growing, heralding the onset of the CO<sub>2</sub> assisted glacial.

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# An Approach to Change Detection

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## Abstract

Statistical methods for detecting change of parameters at unknown time are discussed and illustrated for univariate time series. An approach to boundary detection for spatial data is proposed and illustrated.

## Introduction

Although concern exists over many environmental issues including acid deposition, unsafe disposal of nuclear and other toxic waste, and health problems caused by air pollution and contamination of the water supply, the environmental issue perhaps of most concern is that of climate change.

Monitoring or surveillance of a time series of observations with a view to detecting trends or changes in parameters is an important statistical tool that finds, or can find, application in the analysis of data relating to each of the environmental issues listed above. For example, is global warming occurring? Various analyses of temperature data have suggested an increase of half to a third of a degree Celsius over the past hundred years. However, questions regarding the comparability over time of the data raise doubts about the conclusion. The best monitoring methods must be applied in the future to this time series to ascertain whether any real or continued temperature change occurs. As another example consider the southerly expansion of the Sahara. Is this a long-term trend? Recently acquired satellite data suggest the desert is in retreat. Careful monitoring and analysis will be required to prevent repetitions of the great hardship suffered in the concerned areas. Other examples of important uses of monitoring methods include detection of leakage from toxic waste sites, detection of serious declines in the sizes of populations of various species, and detection of changes in fertility rates in developing countries.

Changes in extent of snow cover and thickness of lake ice are variables expected to be highly correlated with changes in the mean and variance of global temperature. Hence, monitoring of these variables with a view to detecting changes is expected to produce early evidence of more general climatic change.

In each of the above cases the time series data gathered for monitoring purposes are highly stochastic. Hence any useful analysis will require the application of appropriate statistical methodology.

## The Change-point Problem

The statistical problem underlying monitoring or surveillance is the change-point problem. This problem is posed in terms of various statistical models for time sequenced data. Such models are generally characterized by several unknown parameters. These

parameters may change over time, and if the changes, when they occur, do so unannounced and at unknown time points, then the associated inferential problem is referred to as the change-point problem.

The change-point problem gives rise to non-standard inferential problems. These include: detection of changes, estimation of change-points, and estimation of the parameters before and after the change-points. In the case of possible global warming, the sequence of observations to be monitored might be the estimates of annual global mean temperature as compiled by Hansen and Lebedeff (1987). The parameters of interest are the mean and variance of the observations. An hypothesis to be tested could be that there has been no change in global mean temperature. If one assumes there is no particular event that could have triggered a change, then one must ascertain whether the data are compatible with the hypothesis of no change versus that of change at unknown and unspecified time point. If a change is detected then one will wish to estimate the point of change in the parameter (in this case, the mean), and also estimate the values of the parameter before and after the change.

It should be pointed out that the nature of model could also change. In the case of annual global mean temperature, the change might be from that of a constant mean to that of an upward linear trend. The problems of detecting a change and of estimating the change-point still remain, albeit somewhat more complicated.

As noted above, a non-statistical problem associated with the detection of non-stationarities in time series involves the comparability of data collected over a long period of time. Over the course of a century there will be substantial changes in measurement technology, there may be changes in the positioning of gauges, or the local environment for the gauges may be substantially altered. Ascertaining whether a detected change is genuine or due to corruption of the data is a problem that requires for its solution more than the statistical methods associated with the change-point problem.

More complicated change-point problems arise when one considers spatial data. For example, consider the variable that consists of "maximum extent of snow cover", say, in the northern hemisphere. This might be roughly modelled by a circum-global boundary line. This boundary of maximum extent may be further south in some years than in others, or it may have advanced further south in some regions and receded further north in others. A change detection problem may be associated with determining in any given year where the boundary lies using remotely sensed data. Another change detection problem involves detection of whether the boundary has receded, advanced or remained much the same over an extended period of years.

### **The Change-Point Problem for Means**

We now examine some of the statistical methodology developed to deal with the change-point problem.

We first consider a simple statistical model composed of a mean value with noise superimposed upon it. Hence the observation at time  $t$ , denoted by  $y_t$ , is represented as follows:

$$Y_t = b_{0t} + \epsilon(t), \quad t = 1, 2, \dots, n$$

where  $\epsilon(t)$  is one of a set of independent random variables with zero mean and variance equal to  $\sigma^2 > 0$ . The parameter of interest is  $b_{0t}$ . We wish to test the hypothesis  $H_0 : b_{0t} = b_0, t = 1, 2, \dots, n$  against certain alternatives involving changes in the parameter occurring at unknown times.

If we let components of a prior distribution on the change-point be denoted by  $\{P_k\}_{k=1}^{n-1}$  and if we denote the known mean and variance of  $Y(i)$  under  $H_0$  by  $b_0$  and  $\sigma^2$ , then the statistic for testing against one-sided alternatives is

$$C_n^{(1)}(P) = \sum_{k=1}^{n-1} P_k \left\{ \sum_{j=k+1}^n \frac{Y(j) - b_0}{\sigma} \right\}$$

This statistic was derived by Chernoff and Zacks (1964). The comparable statistic for testing against two-sided alternatives, discussed by MacNeill (1974), is

$$Q_n^{(1)}(P) = \sum_{k=1}^{n-1} P_k \left\{ \sum_{j=k+1}^n \frac{Y(j) - b_0}{\sigma} \right\}^2$$

If one restricts attention to the case of "at most one change" (AMOC) with uniform prior, then  $P_k = 1/(n-1)$ .

If  $b_0$  is not known in advance, then it must be estimated by the sample mean. The change-detection statistics for one and two-sided tests become respectively

$$C_n^{(2)}(P) = \frac{1}{\sigma_0} \sum_{k=1}^{n-1} P_k \left\{ \sum_{j=k+1}^n Y(j) - \bar{Y} \right\}$$

and

$$Q_n^{(2)}(P) = \frac{1}{\sigma_0^2} \sum_{k=1}^{n-1} P_k \left\{ \sum_{j=k+1}^n Y(j) - \bar{Y} \right\}^2$$

In the uniform AMOC case, under  $H_0$ ,  $(n-1)C_n^{(1)}$  is normally distributed with zero mean and variance  $(1/6)n(n-1)(2n-1)$ , and  $(n-1)C_n^{(2)}$  is normally distributed with zero mean and variance  $(1/12)n(n^2-1)$ . Under the same assumption  $Q^{(1)}$  and  $Q^{(2)}$  are quadratic forms distributed as weighted sums of  $\chi^2$  variables. Asymptotic distribution theory is given for  $Q^{(1)}$  by MacNeill (1974) and for  $Q^{(2)}$  by MacNeill (1978).

## Parameter Changes at Unknown Times in Regression Models

Test statistics discussed in the previous section were directed at detecting changing mean levels. To accommodate similar detection problems when data follow non-constant trends it is necessary to consider models whereby data may be generated by different regression regimes in different parts of the time domain.

We consider the linear regression model

$$Y = X\beta + \epsilon \quad (1)$$

where  $\epsilon' = (\epsilon_1, \epsilon_2, \dots, \epsilon_n) \sim N(\mathbf{0}, \sigma^2 \mathbf{I})$ ,  $\beta' = (\beta_0, \beta_1, \dots, \beta_{p-1})$  is the parameter vector,  $Y' = (Y_1, Y_2, \dots, Y_n)$  is the vector of dependent observations and  $X$  is the design matrix with

$$X = \begin{bmatrix} 1 & X_{11} & X_{12} & \cdots & X_{1,p-1} \\ 1 & X_{21} & X_{22} & \cdots & X_{2,p-1} \\ \vdots & \vdots & \vdots & & \vdots \\ 1 & X_{n1} & X_{n2} & \cdots & X_{n,p-1} \end{bmatrix}$$

We regard (1) with  $\beta$  unspecified as the model describing the null hypothesis. Alternative hypotheses are specified by letting  $\beta_i$  be the parameter vector for the  $i$ th observation;  $\beta_i$  may vary with time. The statistic proposed by MacNeill (1982) is

$$Q_n^{(3)} = \sigma^{-2} \sum_{k=1}^{n-1} R' X_{(k)} X'_{(k)} R$$

where  $R$  is the vector of raw regression residuals obtained by maximum likelihood fit and where  $X_{(k)}$  is the design matrix  $X$  with the first  $k$  rows set equal to zero. Since distribution theory for  $Q_n^{(3)}$  is complex, an alternative statistic was proposed by MacNeill (1978) for polynomial regression. This statistic is

$$Q_n^{(4)} = \sigma^{-2} \sum_{k=1}^{(n-1)} (R' \mathbf{1}_{(k)})^2$$

where  $\mathbf{1}'_{(k)} = (0, 0, \dots, 1, 1, \dots, 1)$ , i.e., it is the unit vector with the first  $k$  components set equal to zero. Distributional results for  $Q_n^{(3)}$  are given by Jandhyala and MacNeill (1989, 1991) as is additional distributional theory.

### Estimating the Point of Change

We make the following assumptions regarding the mean of the observations:

$$\begin{aligned} b_{0t} &= b_0 & t &= 1, 2, \dots, m \\ &= b_0 + \delta & t &= m + 1, \dots, n \end{aligned}$$

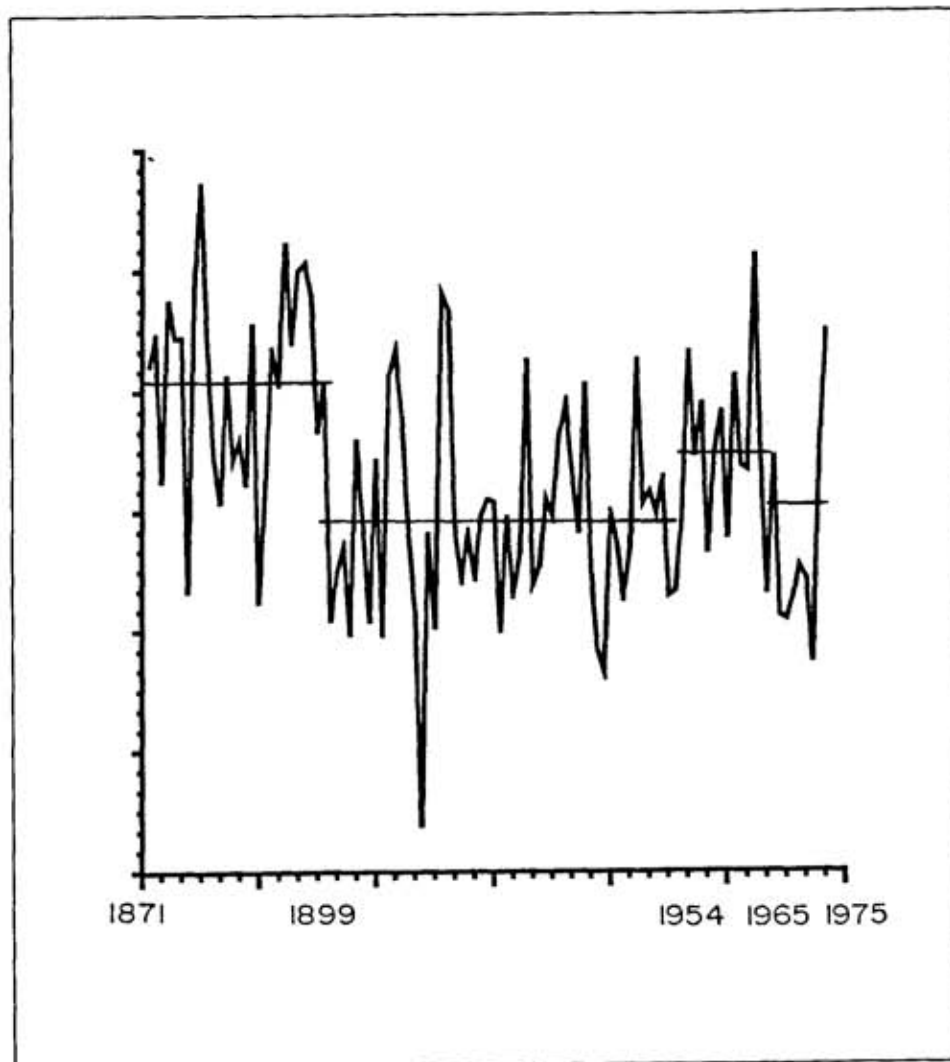
In other words, a change in the mean has taken place at the time point  $m$ . We define the following statistics:

$$\bar{Y}_m = \frac{1}{m} \sum_{i=1}^m Y_i ,$$

$$\bar{Y}_{n-m} = \frac{1}{n-m} \sum_{i=m+1}^n Y_i ,$$

$$S_m^2 = \sum_{i=1}^m \{Y(i) - \bar{Y}_m\}^2 ,$$

$$S_{n-m}^2 = \sum_{i=m+1}^n \{Y(i) - \bar{Y}_{n-m}\}^2$$



**Figure 1.** Change-Points for Nile discharges at Aswan.



Then estimate as the change-point  $m^*$ , that value of  $m$  that minimizes,

$$S_{nm}^2 = S_m^2 + S_{n-m}^2 .$$

That is:

$$S_{nm^*}^2 = \min_m S_{nm}^2 .$$

This methodology has been applied to the annual Nile discharges at Aswan by MacNeill, Tang and Jandhyala (1991) with the results shown in Figure 1.

### Change detection for Spatial Data

The methodology discussed above can be generalized for use in detecting boundaries in spatial data. For example, consider a two-dimensional array of observations:

$$Y_{ij} = b_{0ij} + \epsilon_{ij}, \quad i = 1, \dots, n_1, \quad j = 1, \dots, n_2,$$

with  $\epsilon_{ij} \sim N(0, \sigma^2)$ . We will assume the existence of a boundary in this 2-dimensional grid on one side of which  $b_{0ij} = b_0$  and on the other,  $b_{0ij} = b_0 + \delta$ . The problem is to estimate the location of the boundary. A statistic for estimating the boundary's location is defined as follows. Let  $B$  be a collection of points,  $(i, j)$ , on the grid and let  $\bar{B}$  be the complementary set. Let  $n_B$  be the number of grid points in  $B$  and  $n_{\bar{B}}$  be the number of grid points in  $\bar{B}$ ;  $n_B + n_{\bar{B}} = n_1 n_2$ . Let:

$$\bar{Y}_B = \frac{1}{n_B} \sum_{(i,j) \in B} Y_{ij} ,$$

$$\bar{Y}_{\bar{B}} = \frac{1}{n_{\bar{B}}} \sum_{(i,j) \in \bar{B}} Y_{ij} ,$$

$$S_B^2 = \sum_{(i,j) \in B} \{Y_{i,j} - \bar{Y}_B\}^2 ,$$

$$S_{\bar{B}}^2 = \sum_{(i,j) \in \bar{B}} \{Y_{i,j} - \bar{Y}_{\bar{B}}\}^2 .$$

Then estimate as the boundary  $B^*$ , that boundary  $B$  that minimizes,

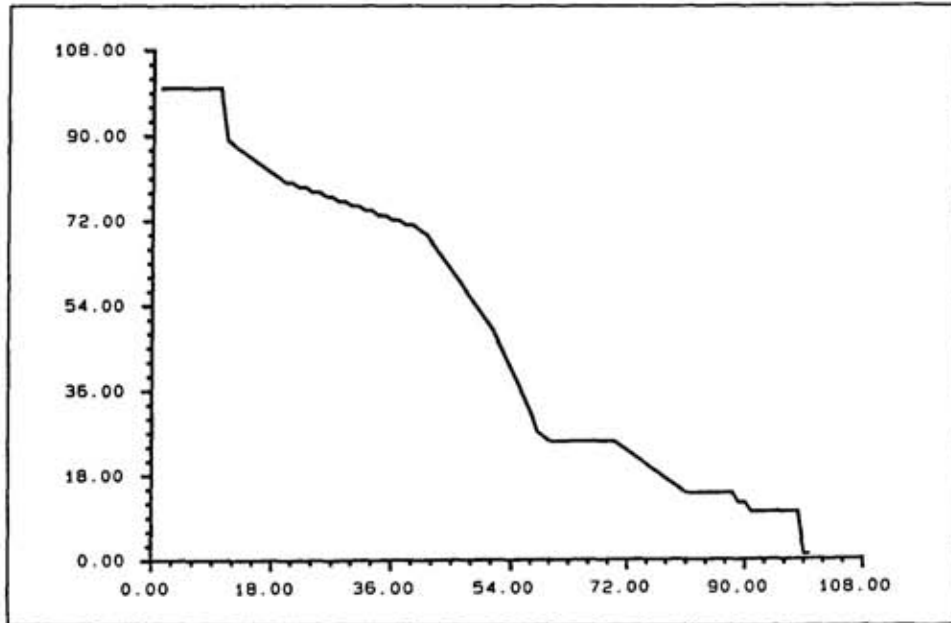
$$S_{B^*B^*}^2 = S_B^2 + S_{\bar{B}}^2 .$$

That is:

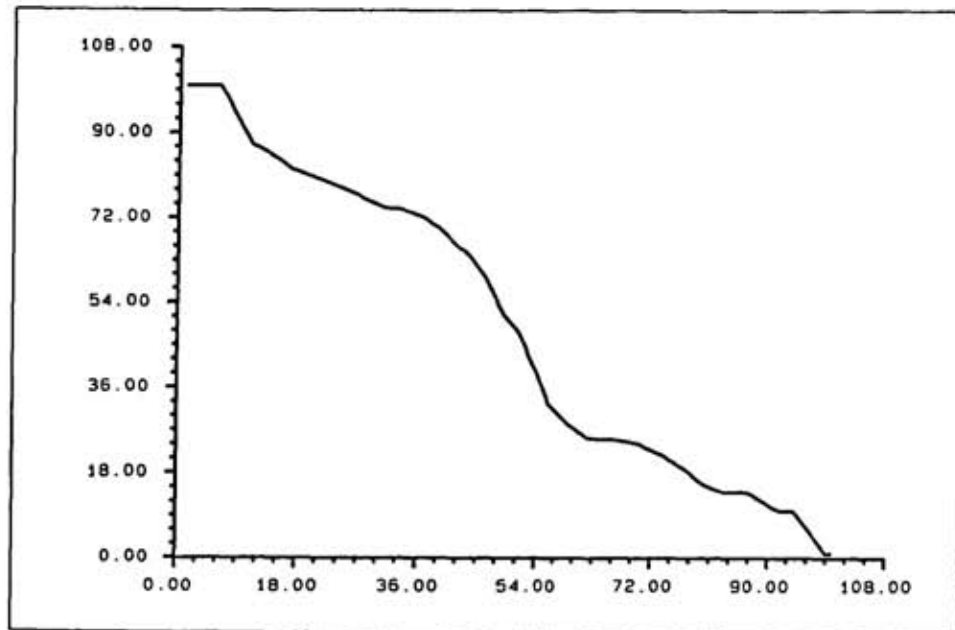
$$S_{B^*B^*}^2 = \min_B S_{BB}^2 .$$

For even moderate sized  $n_1, n_2$  the number of possible boundaries becomes unmanageably large. To reduce the number of boundaries to be considered in the minimization process to a more manageable size, one can make assumptions about

continuity and smoothness of the boundary, and one can use the uni-dimensional methods discussed in previous sections. By this we mean, fix  $i$  between 1 and  $n_1$  and carry out a change-detection and, if required, a change-point estimation procedure on  $Y_{ij}$  ( $j = 1, \dots, n_2$ ). The collection of change-points so obtained will then provide a first estimate of the location of the boundary. Then, minimization of  $S_{BB}^2$  can be carried out on contiguous boundaries with the expectation that one will have eliminated most of the unlikely candidates.



**Figure 2.** Boundary in a 100 X 100 region.

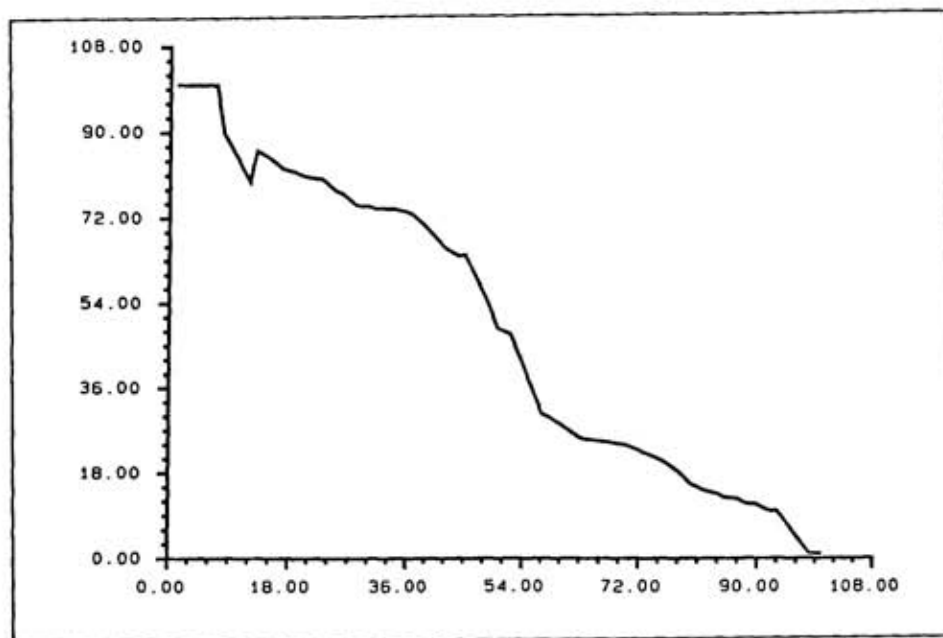


**Figure 3.** Estimated boundary for a mean change of  $2\sigma$ 's.

If one is prepared to make continuity and smoothness assumptions on the boundary, then one may smooth the estimated boundary, for example by using moving averages.

Since small changes in mean are harder to detect than large changes, and since more observations make change detection easier, it is possible to sum sets of  $2k + 1$  contiguous columns of observations and treat them as though they were one column. If the boundary were horizontal then the test would possess the optimality properties of the uni-dimensional test.

As an illustration of this methodology, we consider a grid with  $n_1 = n_2 = 100$ . Then if  $b_0 = 0$ ,  $\delta = 2$ ,  $\sigma^2 = 1$ , and the boundary is as illustrated in Figure 2, we find that application of the methodology with  $k = 3$  and with 5-point smoothing yields a good estimate of the location of the boundary; see Figure 3. If the size of the change is reduced to  $\delta = 1$ , then the methodology has greater difficulty in precisely estimating the boundary, particularly when  $k = 1$ . However if  $k = 3$ , the estimate is as given in Figure 4.



**Figure 4.** Estimated boundary for a mean change of one  $\sigma$ .

The more oblique is the boundary to one of the axes, the greater is the difficulty in determining the boundary. However, a rotation of the axis of direction of the univariate testing procedure will reduce this effect. Of course, the optimal direction will typically not be specified in advance.

It should also be noted that difficulty in identifying changes in parameters increases as the change-point approaches an edge of the space under consideration.

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# A Southern Hemisphere Snow Cover Climatology

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## Introduction

As noted by Matson, et al. (1979), snow cover is a significant climatic index and reflects the dynamic nature of climate. Previously, Namias (1960) noted that large variations in winter snowfall (also snow cover extent) are manifestations of short period climate fluctuations, just as periods of drought, flood, and increased hurricane activity reflect fluctuations in climate. Matson, et al. (1979) further noted that, until 1966, the monitoring of the areal extent of continental snow cover was limited to point measurements and extrapolations between these points. Mountainous and sparsely populated areas were poorly represented in this type of monitoring. It is significant to note that since October 1966, Northern Hemisphere continental snow cover extent has been monitored on a weekly basis by means of satellite data. These weekly charts for the period November 1966 through December 1981 were rendered into a digitized format and since then are being digitized by the National Earth Satellite Data Information Service (NESDIS) of the National Oceanic and Atmospheric Administration (NOAA) in the United States.

As the Northern Hemisphere archive continues to grow, it is not surprising to find more and more researchers extolling the value of these data as an "index" for climate monitoring. Although the necessary satellite imagery exists, a similar "index" has never been prepared for the Southern Hemisphere. In fact, very little is known about the spatial and temporal variability of snow cover in the Southern Hemisphere. Based on Lockwood (1979) and Lamb (1972), the primary areas of Southern Hemisphere snow cover (and therefore, areas suitable for satellite monitoring) are New Zealand, southeastern Australia, the highland areas in the southern tip of Africa and South America.

Therefore, it is the purpose of this study to create a Southern Hemisphere snow cover data archive which is compatible in gridded resolution and identical in temporal extent with the already existing Northern Hemisphere snow cover archive. This report provides an over-view of the creation of this Southern Hemisphere snow cover archive, presents some preliminary findings and outlines the research potential of this unique data set.

## Southern Hemisphere Snowfall Meteorology

Snowfalls over southern Australia are primarily associated with a strong southerly air flow, which is likely to occur when a deep mid-latitude depression is immediately followed by a very large anticyclone. Such synoptic systems can produce snowfalls over large areas of southeastern Australia. Heavy mountain snowfalls can occur even during the summer. Gentilli (1971) reported the occurrence of a summer snowfall that resulted from a deep trough formed by the former tropical cyclone "Audrey" and a very large depression south of Tasmania. This January 14, 1964 storm produced enough snow on Mt. Kosciasko (65km from Canberra) to trap 740 cattle for 17 days. Winter snow in New Zealand generally results from disturbances in the southwesterly airflow. A series of cold fronts, generally associated with a very deep depression passing to the south of New Zealand,



moves from the southwest to the northeast over the country at intervals of 12 to 36 hours. The last cold front in a series usually brings snow to low altitude locations in the Southland and the higher altitude locations in many other areas. During the winter, the snowline usually descends from about 2,500 meters to 1,000 meters on North Island and from about 2,000 meters to 800 meters on South Island (Gentili, 1971). Snowfalls are most frequent in July and August, but may occur at higher elevations at any time of the year. Snow falls over the southern tip of Africa occur primarily in the few highland locations and are the result of exceptionally strong depressions travelling far enough north to advect cold air masses into South Africa. Although, rare, snowfalls have occurred in the lower elevations in Cape Town, South Africa.

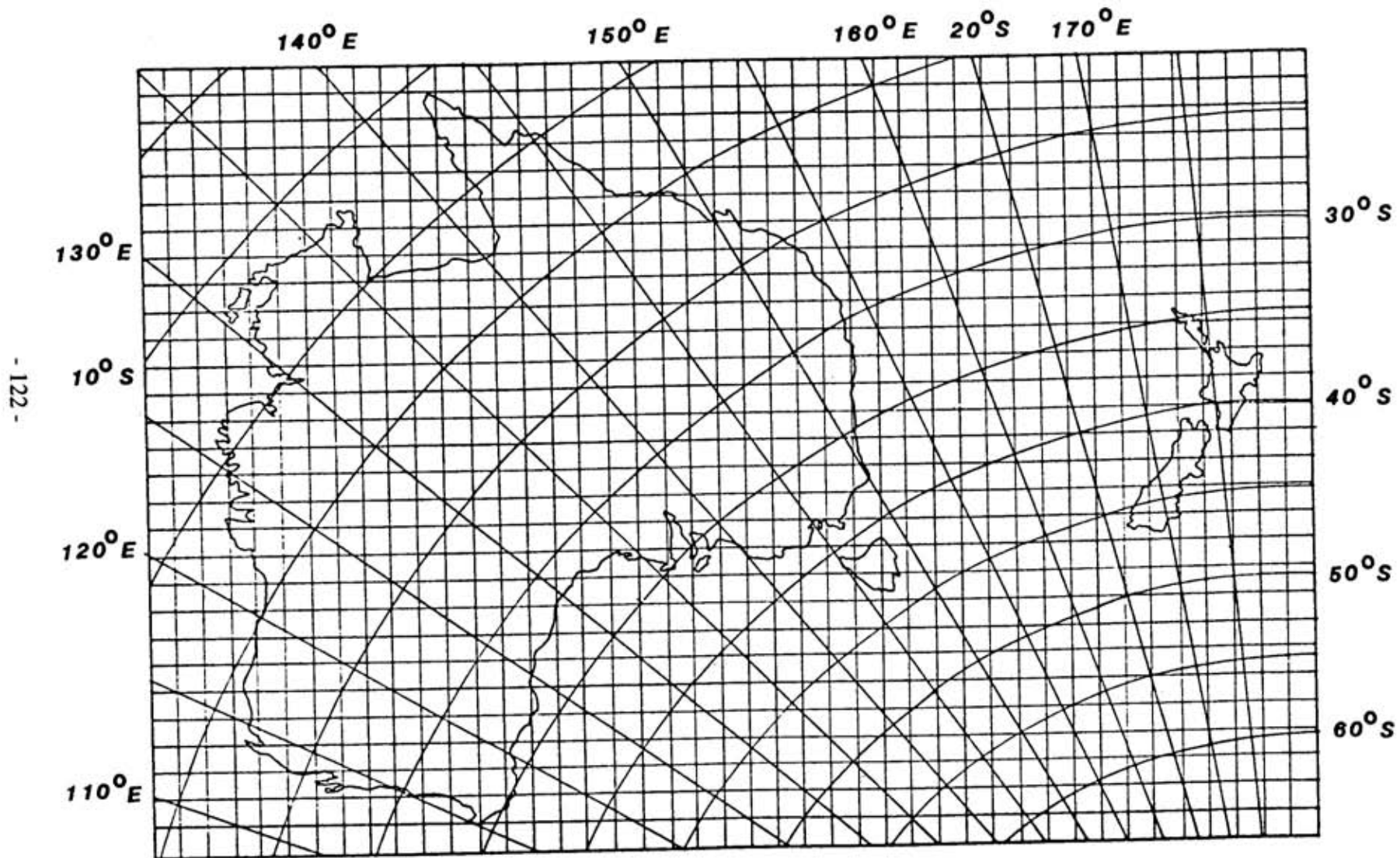
## **Analysis Procedures**

Defense Military Satellite Imagery (DMSP) were utilized in this study. The imagery were provided from a polar orbiting satellite which provides pole to pole swaths covering the entire study area at least once a day. These imagery were then analyzed and the snow cover boundaries were drawn for each day. During time periods of obscuring cloud cover, the last visible boundary for the affected regions was carried over onto the next map. The map projection utilized was a Polar Stereographic projection, identical to the map projection utilized in the creation of the Northern Hemisphere snow cover maps (Dewey and Heim, 1981). Using the exact weekly dates from the Northern Hemisphere analysis, the Southern Hemisphere charts were grouped into weekly (7-day) average maps. This procedure is previously outlined in a NOAA Technical Report NESS 87 (Dewey and Heim, 1981).

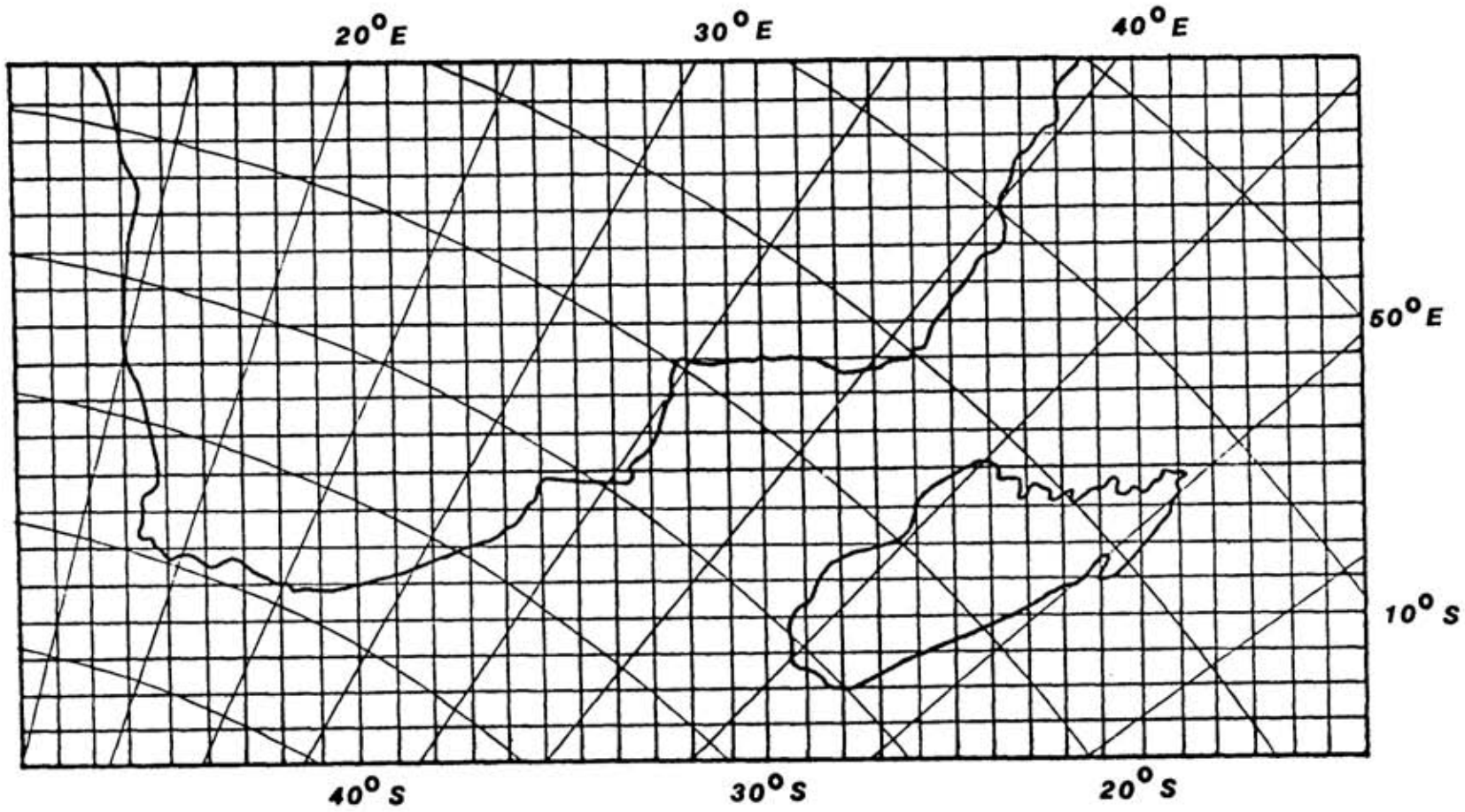
The weekly charts were then hand-digitized using the identical grid system that was utilized for the Northern Hemisphere study (which will allow for inter-hemispheric comparisons). This 89 x 89 grid is the same grid which is utilized in the NOAA/NWS PE weather forecast model and the satellite based earth radiation budget atlas of Winston, et al., (1979). Any grid box that was 50% or more covered by snow was considered to be completely covered and conversely, any grid cell with less than 50% snow cover was considered "open". Figures 1, 2, and 3 illustrate this grid for the three major regions of potential snow cover (Australia-New Zealand, Southern Africa, and South America). Figure 4 illustrates the problem of grid resolution for snow cover in the Australia-New Zealand region. This figure represents one of the more extensive snow covers in this area. The problem of sub-grid resolution for snow cover in Africa, although not shown here, was quite similar to Australia-New Zealand. The decision was then made to map only the South American snow cover for this study. The sub grid size snow covers for these excluded areas could in the future be analyzed using any finer grid system. Figure 5 illustrates the extensive nature of the South American snow cover for the peak week of snow cover in 1984. The region of South America between 0° and 15° S latitude was excluded from the analysis due to long periods of persistent orographic and convective cloudiness obscuring the snow fields on the Andes Mountains. Snow cover data in South America south of this region were found to be reliable. Snow cover frequencies and interannual variations of snow cover in this region should be significant.

## **Analysis Of The Digital Archive**

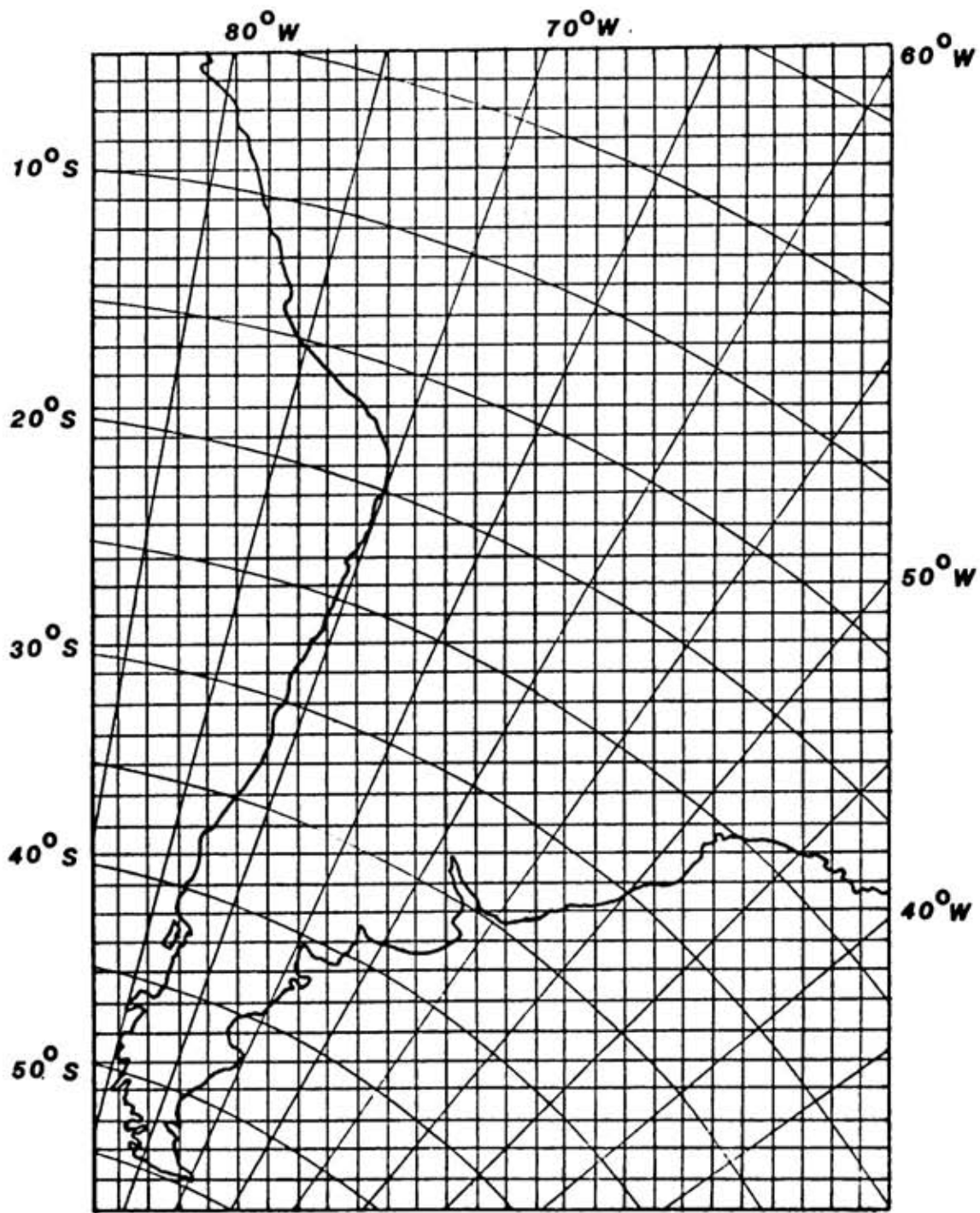
The current data archive covers the time period from 1974 through 1986 and will be expanded during the coming months to extend from 1967 to present (duplicating the same time period as the Northern Hemisphere snow cover archive). The computerized digital archive of South American weekly snow cover charts was utilized to prepare several digital



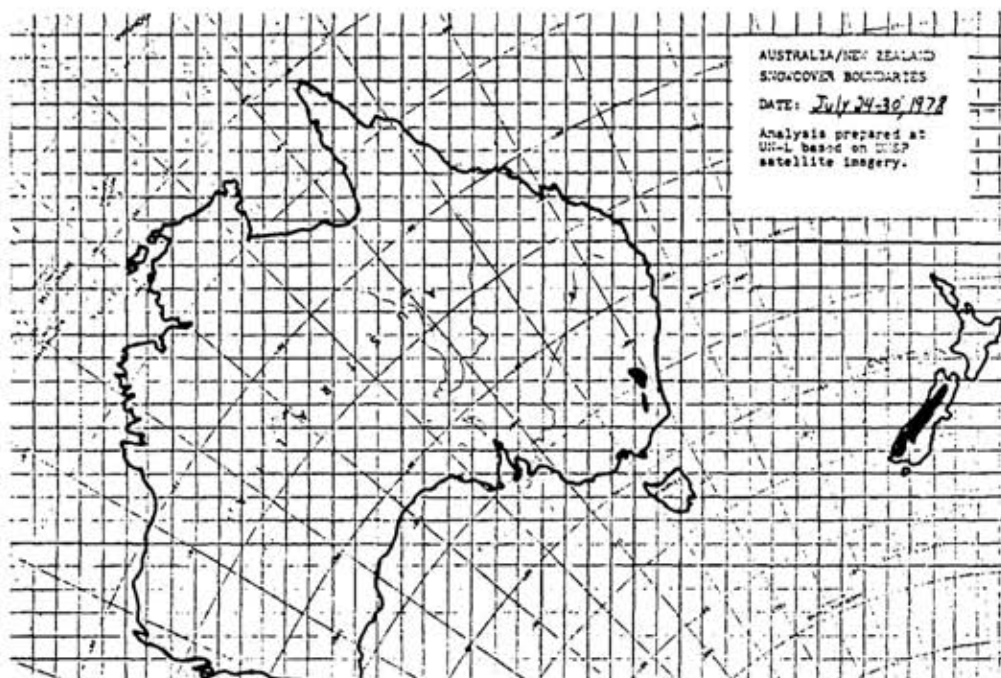
**Figure 1.** Analysis grid for Australia-New Zealand region with the attendant digitizing grid.



**Figure 2.** Analysis grid for Southern Africa with the attendant digitizing grid.

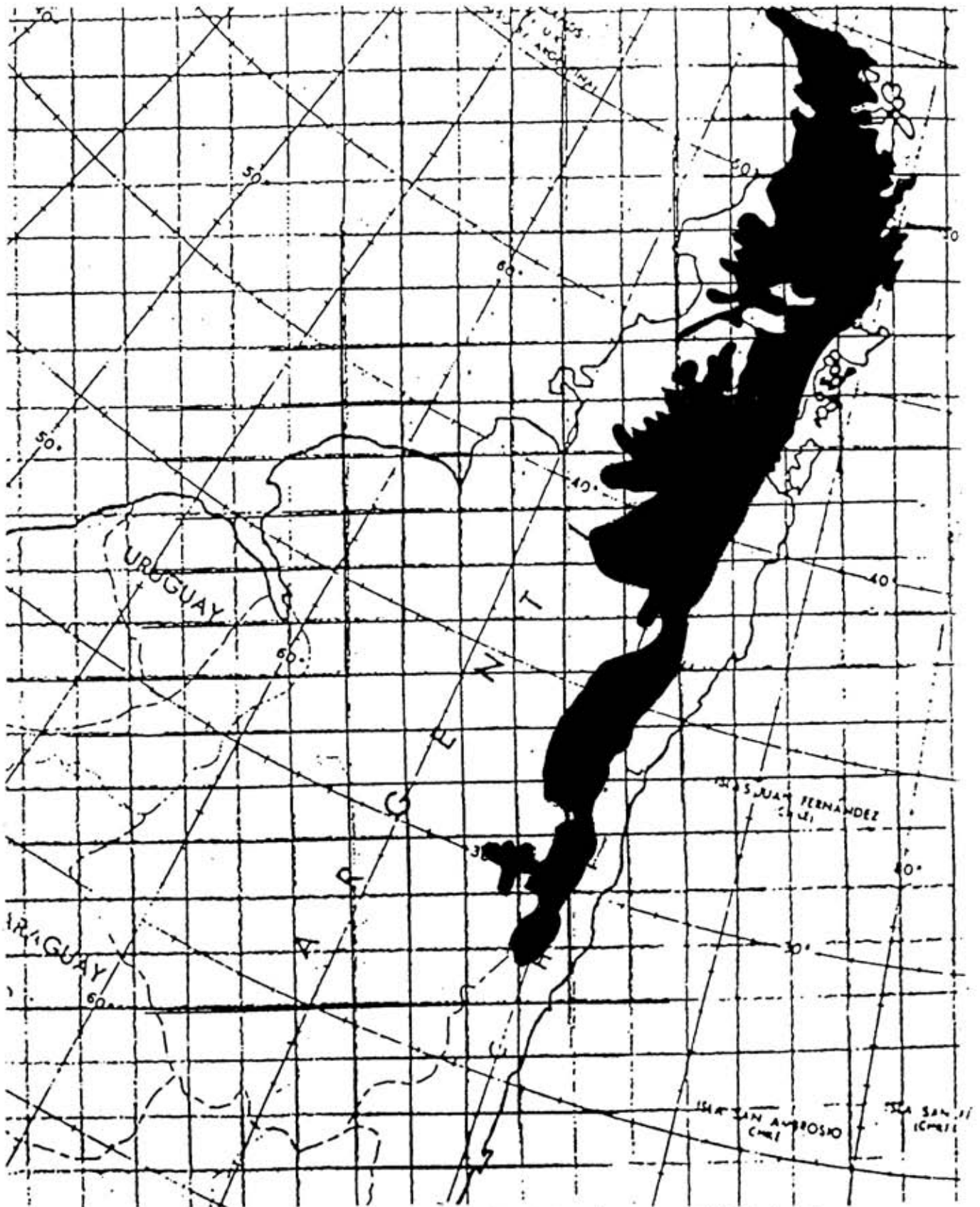


**Figure 3.** Analysis grid for South America with the attendant digitizing grid.



**Figure 4.** Snow cover analysis for one week in Australia-New Zealand snow cover for the week of July 24-30





**Figure 5.** Illustrates the extensive nature of the South American snow fall during the week of peak snow cover in 1984

snow cover products. Using the grid box area matrix, as determined in the Northern Hemisphere study, weekly snow cover areas (in square kilometers) were computed for South America. These weekly areas were averaged to determine the weekly mean for the study period and the annual mean for each year as well as a time series of the annual means. Figure 6 illustrates the variation in weekly snow cover area for 1978 in South America. As can be dramatically seen, the week to week variation in snow cover shows a high degree of variability. This variability is not as dramatic in the Northern Hemisphere climatology (Dewey, 1987). Although not illustrated here, time series were created for each of the other years and time series for each of the weeks (week #18, #19, #20 etc., 1974-1986) were also created. It is interesting to note the range in peak weekly snow cover for South America:

### **RANGE IN PEAK EXTENT OF SOUTH AMERICAN SNOW COVER**

**MOST: WEEK #30, 1984:  $12.01 \times 10^5 \text{ KM}^2$**   
**LEAST: WEEK #31, 1979:  $6.92 \times 10^5 \text{ KM}^2$**

The time series of the annual mean snow cover for South America for the time period 1974-1986 is presented in Figure 7.

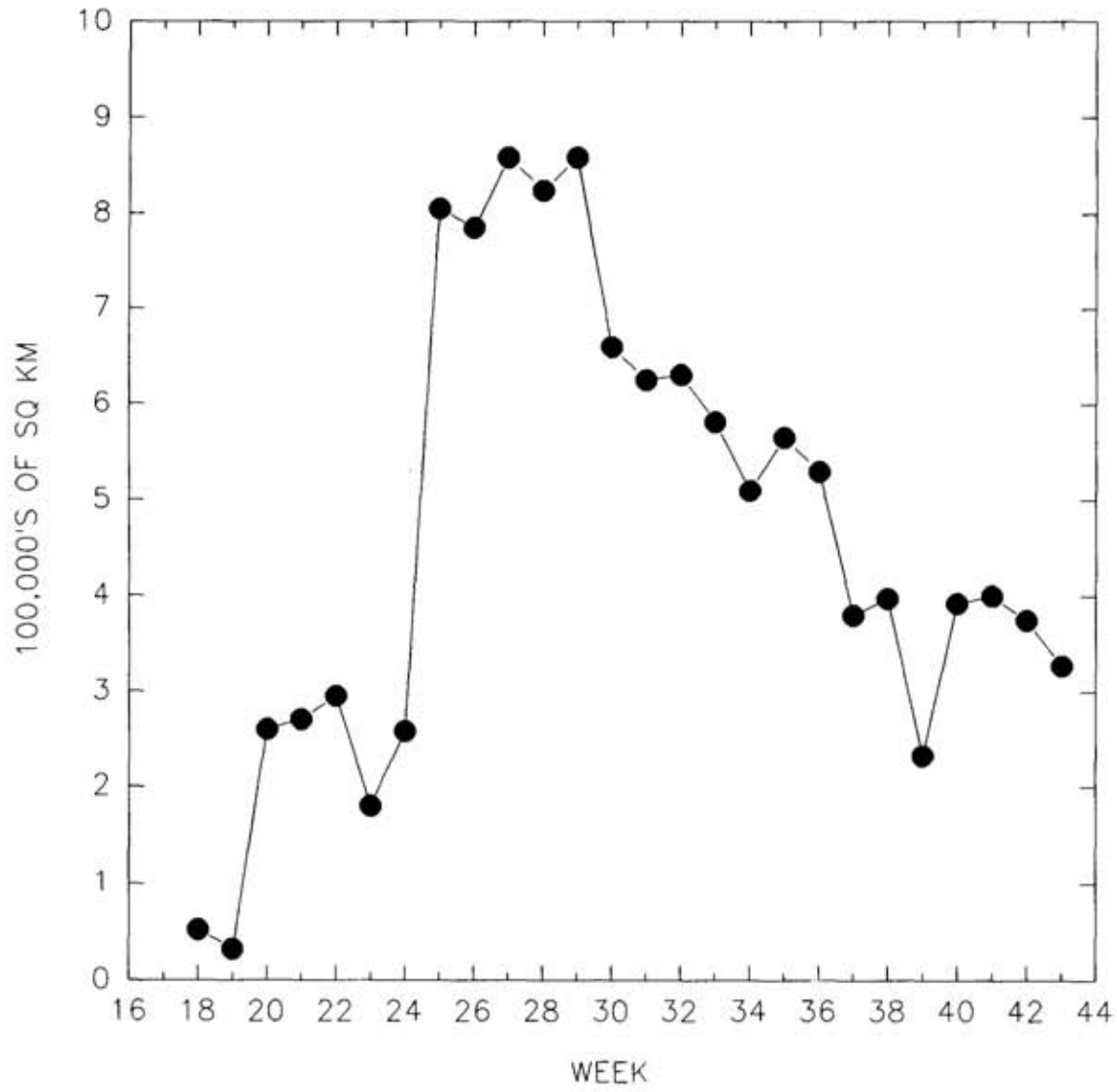
The figures for 1984, 1985 and 1986 should be considered "first estimates" and have not been verified at the time of the preparation of this report. A preliminary investigation of a relationship between South American Snow cover (using this data archive) and the Southern Oscillation was also conducted (Cerveny, Skeeter, and Dewey, 1987). South American snow cover area during the May to October portion of the snow season was found to be correlated significantly with the winter Southern Oscillation Index (SOI) values of the same year. The relationship is inverse with periods of low SOI values being associated with extensive South American snow cover while high SOI values tended to occur during years of diminished snow cover. An equatorward movement of the Southern Hemispheric westerlies during the winters experiencing low SOI values would result in greater precipitation rates between  $20^{\circ}$  and  $40^{\circ}$  S, due to increased cyclonic passage. The increased cloud cover which produces cooler temperatures combined with increased precipitation would most likely result in increased snow fall for this area. This preliminary study provides the first application of satellite snow cover imagery to the investigation of the Southern Oscillation. A longer Southern Hemisphere satellite record would obviously permit a more detailed investigation.

### **Conclusions**

Caution must be utilized in drawing any sweeping conclusions or generalizations about these figures. However, it is apparent that the extent of snow cover in this portion of the Southern Hemisphere is significant and quite variable from year to year.

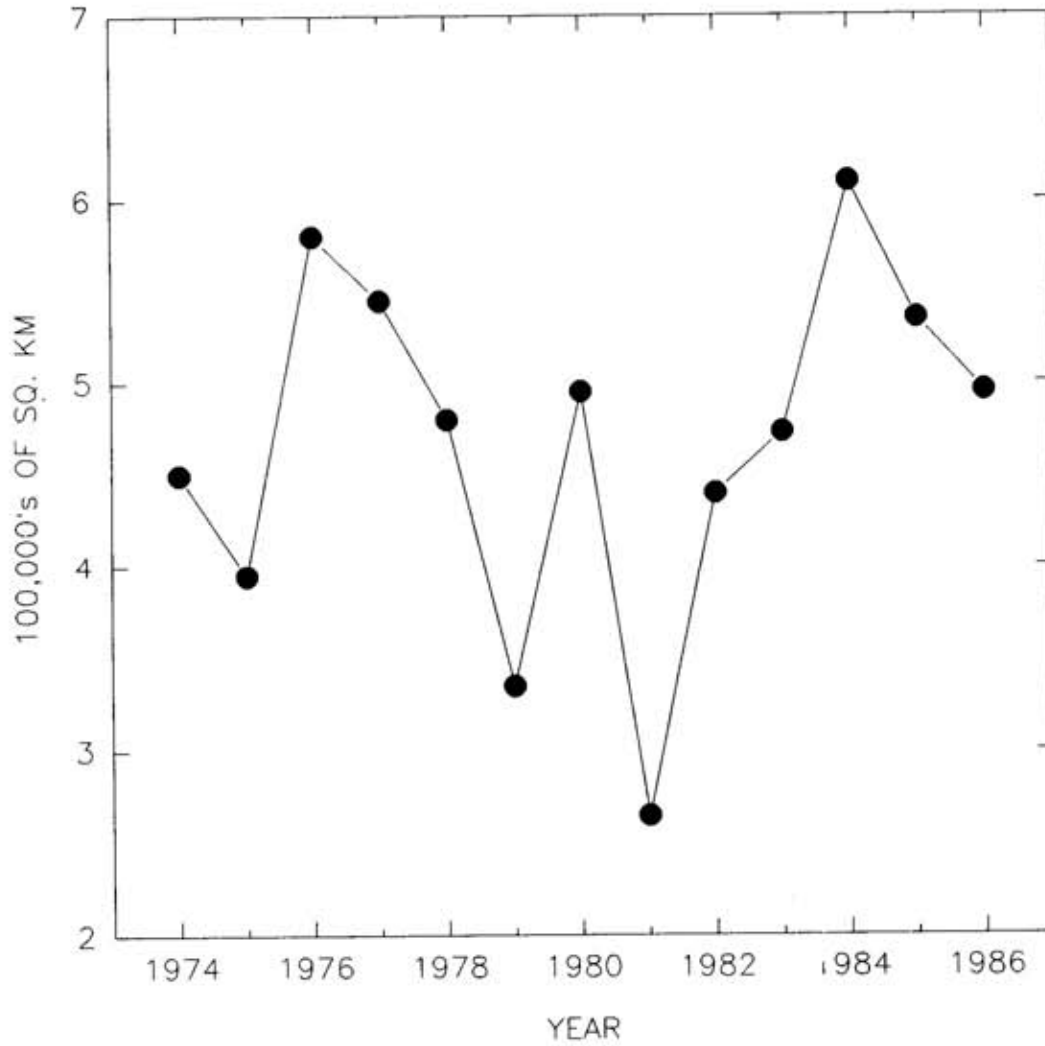
When this data archived is extended temporally to equal the Northern Hemisphere record, inter hemispheric comparisons of climate variability and trends can then be conducted. The Southern Hemisphere snow cover archive will also serve as a useful climate "index" for the southern hemisphere and perhaps will lead to a better understanding of how each hemisphere is responding to the well publicized "global greenhouse warming" phenomenon. This data archive should also provide useful service for atmospheric circulation modellers desiring snow cover information for the Southern Hemisphere. The

### SOUTH AMERICA SNOW COVER - 1978



**Figure 6.** Illustrates the variation in weekly snow cover area for 1978 in South America

### MEAN ANNUAL SOUTH AMERICAN SNOW COVER



**Figure 7.** The time series of the annual mean snow cover for South America.

Southern Oscillation, particularly as evidenced in ENSO events, has had a dramatic impact on the global climate. The existence of a relationship between South American snow cover and the Southern Oscillation may prove beneficial in our understanding of the spatial and temporal mechanisms of the climate system. Finally, it is hoped that as this unique data archive for the Southern Hemisphere continues to grow in temporal extent, so will the potential number of uses for this data archive grow within the research community.

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# **A Brief Overview on Snowcover/Climate Research at Lanzhou Institute of Glaciology and Geocryology (LIGG) Chinese Academy of Sciences**

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## **Avalanche and Snowcover Physics Study in Chinese Western Tianshan Mountains, During 1965-1975.**

- Highway 0503 across Tianshan from north to south was frequently closed in winter and early spring because of 60-110 cm snow on the ground in normal years and especially of avalanche and wind-drift snow hazards.
- A group of scientists from the institute started avalanche research in this area. Soil dams in avalanche valleys, channels and wind fences were built up to protect the highway from the hazard. Snowpack temperature, snow density, grain size, free water content and metamorphism were measured and observed by simple instruments near the base camp in the research area.
- Research papers and 3 books had been published in late 70s and 80s. An avalanche station was set up in that area, which is now operated by Xinjiang Institute of Geography. Energy balance and snow melt were studied in recent years.

## **Remote Sensing And Snowmelt Runoff Forecasting In Qilian Mountains And In The Upper Reaches Of Yellow River**

- 25% surface water in Hexi region comes from glaciers and snowcover in Qilian mountains; 12% of the annual total from the snowmelt in spring.
- NOAA/TIROS APT and AVHRR CCT data were used to monitor the snowcover area in the mountains by the NAC-4200f multicolor data system in the institute during 1980-1991.
- Snowcover percentage, daily precipitation and air temperature at several ground stations were the factors for spring runoff forecast in an exponential equation. Good forecast was provided to the agriculture and hydro-power operation in the upper reaches of Yellow River on Tibet Plateau.
- A project on monitoring and protecting heavy snow disaster in inner Mongolia pastures is carried out by remote sensing and ground observation for years in the "international decade of nature disaster reduction".
- A joint study of NASA and LIGG on microwave remote sensing of snowcover in west Chinese mountains is ongoing.

## Snow Hydrology Study at Tianshan Glaciological Station

- Tianshan Glaciological Station  
43.06N, 86.50E; 3545 m a.s.l. Upper streams of Urumqi River

	AIR TEMPERATURE (c)	PRECIPITATION (mm)
ANNUAL	-5.4	430
JANUARY	-15.9	3
JULY	4.7	106

About 30 years mass balance measurement on glacier No.1

- Determination of spatial variation of snow depth, density and S.W.E. on glacier and ice-free slopes of various orientations by means of snow survey of high resolution of 5 - 20 metres;
- Stratigraphy and temperature differences of the snowpack on glacier and ice-free ground;
- Comparison of ground snowcover storage to snow/rain gauge measurement;
- Snow surface sublimation and weak melting during winter;
- Intercomparison measurement of snow density with different instruments, e.g. Samplers (tube/cutter) and balances;
- Glacier/snowmelt runoff regime and simulation in alpine areas;
- Responses of surface runoff to climate change;
- Chemical sampling and analysis of snow and melt water.

### Snow Cover Resource and its Fluctuation

#### 1) Chinese snow data set

- 2300 weather/climatic stations, during 1951-1980
- daily snowfall (mm)
- daily snow depth (cm)
- snow density every 5-10 days ( $\text{kg/m}^3$ ).

#### 2) Basic graphs

- annual snowfall in China
- duration of snowcover in China
- mean and maximum snow depth in China

#### 3) Fluctuation of time series of annual snow mass

- annual snowfall & winter snow depth proportional to global mean temperature;
- heavy snow correlated with ENSO ( $r=.63$ ,  $n=30$ );
- light snow associated with large volcanic eruptions;

4) Tendency of fluctuation

- generally increasing over China, on the average;
- going up in Tibet Plateau (southwestern China) and the low reaches of Yenzi River (southeast China);
- coming down in lowlands and some of the mountains in north China.

### **Problems of Snow Cover/Climate Research in China**

1) Field experiences of snow cover measurement

- only two field stations in northwestern China for snow study, need one or more in northeast mountains of maritime climate;
- old instruments and equipment in field and lab;
- few young snow scientists who have rich experience of snow study.

2) Remote sensing image data

- Chinese satellite can not provide high resolution data for snow cover/climate study in small mountain areas;
- very limited image data from the limited receiving stations for foreign satellites;
- few aerial-photos available recently because of the high cost of specific flight.

3) Scientific communication

- language barrier;
- a few foreign professional journals available in China.

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# An Overview of Research on Snow in Alaska

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Glaciers cover an area of about 102,000 km<sup>2</sup> in Alaska and the adjacent parts of Canada (75% of this area is in Alaska). For perspective 12 states in the U.S. have smaller areas than this ice-covered area which is about equal to that of the Queen Elizabeth Islands of Canada. On planet Earth, they are exceeded in area only by the Greenland ice sheet (1.7 x 10<sup>6</sup> km<sup>2</sup>) and the Antarctic ice sheet 1.3 x 10<sup>7</sup> km<sup>2</sup>. The Alaska-Yukon glacierized area is more than twice the glacierized area in the Himalaya and Karokorum ranges combined (Benson and others, 1986). But, impressive as it is, perennial snow and glaciers cover only 5% of Alaska's total area of 1.5x10<sup>6</sup> km<sup>2</sup>, all of which is subjected to seasonal snow. This paper presents an overview of three decades of research on Alaskan snow.

Snow forms a thin veneer on the earth's surface over most of Alaska for 1/2 to 3/4 of the year. The physical properties of this snow layer and the physical processes which occur in, above, and below it are important and fascinating. The Alaskan snow differs from the hydrologically important mountainous snow of the western United States in that its temperatures are lower, steeper temperature gradients occur in it, and there is less of it per unit area; however, it lasts longer and enters more directly into human activity as snow itself rather than serving primarily as a cold storage water reservoir. This is of course especially true of snow which falls in glacier basins and enters into the complex glacier-hydrology system. Some of it may appear as runoff during the same year it was deposited, but much of it becomes locked in the glacier system for many decades before it appears as runoff.

Indeed, Alaska is virtually a made-to-order snow laboratory because it contains maritime, continental, and Arctic climates in proximity. Striking differences exist in the snow cover from one climatic region to the next. This fortuitous situation is the result of two sharply defined climatic boundaries which cross Alaska:

- (1) The Alaskan coastal ranges separate the north Pacific maritime climate from a severe continental climate.
- (2) The Brooks Range separates the interior continental climate from the Arctic polar basin climate.

Each of the three climatic regions, contains its own characteristic snow cover. Pruitt (1970) distinguished two primary North American snow types, "tundra snow" and "taiga snow," which are widespread in Canada and Alaska. To these Benson (1967, 1969, 1982) added "maritime snow" as a third type and defined the climates and their snow types as follows:

- (1) Arctic. The Arctic Slope north of the Brooks Range has the climate of the Arctic polar basin (Conover, 1960). Its precipitation comes from cyclonic disturbances moving eastward from the Bering Sea or from along the Siberian Arctic coast. About 65 to 80% of it comes as snow, the snow cover lasts for nine months and is wind-packed,



dry, and sastrugi-sculptured. Following Pruitt (1970) we refer to this snow as Tundra snow.

- (2) Interior. The interior, between the Brooks and Alaska Ranges suffers an extreme continental climate. Most of its precipitation is from cyclonic disturbances which move eastward from the Bering Sea. About 30% of it comes as snow, which lasts for slightly more than 6 months. Its most notable physical characteristic is the low-density, loosely consolidated, depth hoar which makes up most of the snowpack in the lowland brush and forest areas (Trabant and Benson, 1972). Following Pruitt (1970) we refer to this snow as Taiga snow.
- (3) Maritime. The coastal mountains and lowlands of southeastern and southcentral Alaska receive heavy precipitation from Pacific cyclonic disturbances which move through the Gulf of Alaska. The maritime snow pack is thick, often exceeding 10 m; it may be wet, especially at low altitudes. Snow temperatures are significantly higher than in tundra or taiga snow. It is common along the Pacific rim northward from Japan and California. We refer to this snow as Maritime snow.
- (4) Transitional. In addition to these three major climatic types, a fourth lies south of the Alaska Range and is transitional between the Interior and Maritime zones. It is called the Transitional zone. Climatic conditions alternate between continental and maritime. In the interior this transitional zone also become apparent as one progresses westward toward the Bering Sea, especially west of Koyukuk (about 158°W) on the Yukon-Kuskokwim delta. Here, the temperatures and winds are higher than farther east and the climate becomes more maritime; many of the snow storms are mixed with rain and the snow cover is characterized by significant amounts of icing with depth hoar at the bottom. The snow cover is transitional between the three major types but no special name is applied to it.

Since 1961 we have been doing research on the snow of interior and arctic Alaska. In the interior much of our attention has been focused on the problem of depth-hoar formation. In the arctic, we have attempted to determine the amount of snowfall, the extent of wind transport and deposition, the structure of the snow, its interaction with vegetation and the mechanisms which govern its melting. In the Wrangell Mountains we have studied the non-melting snow of the dry-snow facies, of the glacier facies spectrum (Benson, 1962), in comparison with examples from the polar regions (Benson, 1967, 1968; Wharton, 1966).

George Clagett of the Soil Conservation Service (SCS), U.S. Department of Agriculture, maintains a network of snow-survey sites in Alaska; on the Arctic Slope this involves the use of gages shielded by the Wyoming Blow-Fence (Clagett and others, 1983). Compared with the long record of snow-courses in the "Remote 48 States" the Alaskan program is young, its first report was issued on 11 April 1961. That first report dealt with ten snow courses, including one at Log Cabin, British Columbia. Its longest period of record was nine years, at the Ptarmigon snow course near Palmer, and it contained the first year of record for six of the ten snow courses. The network expanded rapidly. In 1968 there were 81 points with either snow courses or aerial markers but none was on the Arctic Slope. The first Wyoming gage on the Arctic Slope was built in 1975 at Atqasuk on the Meade River about 65 miles SSW of Barrow. It was initially a research site established by the Geophysical Institute of the University of Alaska. This gage and others built by the Geophysical Institute, CRREL, USFS and SCS have all been turned over to the SCS for long-term maintenance and operation. In 1980, the SCS Alaskan snow survey reports included 150 snow courses and 17 Wyoming gages. In 1991 the network contained 164 snow courses and 15 Wyoming gages. The reports are published in

February, March, April, May and June each year by the SCS (201 E. 9th Ave. Anchorage, Alaska 99501-3687.

Matthew Sturm of USACRREL is currently (for the past three years) maintaining intensive research sites along a "traverse" from the Pacific Coast to the Arctic Coast (Sturm and others, 1991). The traverse includes:

1. Tundra snow

Prudhoe Bay ----- on the Arctic coast,  
Imnavait Creek ---- in the northern foothills of the Brooks Range

2. Taiga snow

Glenn Creek  
Engineering Creek ----(both near Fairbanks)

3. Transitional-Maritime

Tsina River ----- north of Thompson Pass about 30 miles from Valdez,  
Low River ----- south of Thompson Pass about 15 miles from Valdez,  
Valdez (These sites go from dry to wet.)

A site is also maintained at Alyeska, west of Anchorage specifically to study deep snow in the Chugach Mountains.

Measurements at each site include: air temperature, short-wave radiation, wind speed and direction, a sonic depth-sounder (measuring to the snow surface from above), a staff gage to measure snow depth, heat-flow meters at the snow-soil interface and sets of thermistors at the snow-soil interface, vertically through the snow and extending into the soil below and the air above, and horizontally in the snow. These sites have been maintained with data-loggers collecting data hourly through the winters of 1989-90, 1990-91 and 1991-92.

This project has resulted in an unprecedented amount of data, which is well-managed and still being processed on Macintosh Computers. Among the results is a clear documentation of the difference in snow-soil interface temperature throughout the winter from one region to another. The interface temperature remains at 0°C in the maritime snow region, at -3°C to -4°C in Taiga snow and -7°C to -21°C in Tundra snow.

## Interior

An extreme continental climate prevails in the interior of Alaska, near Fairbanks. This includes ideal conditions for the formation of depth-hoar in the seasonal snowpack which is generally less than 1m deep. The snowpack lasts for 150-200 days at temperatures well below freezing (-40°C is common) and has a large and persistent thermal gradient (1°C cm<sup>-1</sup> is common) between the ground below and the air above. The resulting depth-hoar has exceptionally low density (0.19 to 0.20 g cm<sup>-3</sup> compared with 0.28 to 0.30 g cm<sup>-3</sup> in alpine regions) and it sometimes spreads upward through the entire snowpack before the end of winter.

To study the formation of depth-hoar we use an experimental arrangement which allows us to compare adjacent snowpacks with identical histories of deposition, but with very different thermal conditions. This is done by building tables which keep the snowpack from contacting the soil, so there is no heat source from below. Depth hoar forms in the snow on the soil but not in the snow on the table. By comparing the snow density and structural profiles repeatedly through the winter we concluded that convection

was involved in the upward mass transfer within the snow (Trabant and Benson, 1972). The thermal effects of convection were demonstrated by Sturm (1989) in an extensive study aimed at determining the flux gradient of the upward moving water vapor, and the thermal effects produced in the snow.

The fractionation of stable isotopes ( $^{18}\text{O}/^{16}\text{O}$  and D/H) during the formation of depth hoar was studied by sampling the adjacent snow packs from nine years. All sampling was done at the beginning of spring before melting occurred. The upward flux of water vapor extends to the soil below. The flux from soil to overlying snow was measured by placing an impervious plastic sheet on a section of soil just before the first snowfall. An ice layer formed under the plastic sheet, before melting began this ice was removed from a known area and weighed to determine the flux; samples of it were also analyzed for stable isotopes. The water vapor from the soil had isotope values more positive than any of the snow and was easily traced in the snow lying on the soil. Significant fractionation of the isotopes took place during the process of depth hoar formation as determined by comparing snowpacks on the three substrates: soil, table, and plastic (Friedman, and others, 1991).

Much of the above research was done on snowpacks which were as free from horizontal variability as possible. We are now working to determine the extent of variability in the natural snowpack and the way it interacts with vegetation. A significant part of this is the effect of "tree wells" formed around spruce trees (Sturm, 1992).

## Arctic Slope

The winter snow on the Arctic Slope of Alaska lasts for nine months each year, yet little is known about it from a scientific point of view. Measurements made on this snowpack since 1961 show that tundra snow has two distinct facies. One is the thin veneer of depth hoar and wind slab that forms over most flat areas of the tundra. The second is the deeper, denser snow that forms drifts in topographic depressions and in the lee of ridges and bluffs.

The veneer of snow which forms on the tundra resembles the top annual unit of snow on the Greenland Ice sheet. It usually consists of a basal layer of depth hoar overlain by a sequence of harder snow. However, it differs markedly from snow on a glacier or ice sheet in that it interacts in a complex way with the tundra vegetation producing marked "horizontal" variability in textural and thermal properties. In particular, in the autumn when the tundra is still warm enough to cause melting, basal snow layers can develop extreme heterogeneity due to the formation of icy places between tundra tussocks. Tundra snow is also subjected to much stronger temperature gradients and higher upward-directed heat fluxes than the equivalent layer of snow on a glacier or ice sheet. This leads to relatively more depth hoar in the tundra snow pack. Average depth ranges from 10 to 40 cm. Wind-slab densities can be as high as  $0.54 \text{ g cm}^{-3}$ , while the depth hoar density is frequently less than  $0.20 \text{ g cm}^{-3}$ . The overall density of the tundra snowpack is slightly less than  $0.30 \text{ g cm}^{-3}$ ; the comparable value for drifted snow generally exceeds  $0.40 \text{ g cm}^{-3}$ .

The structure of snow on the tundra has been studied by excavating trenches and measuring the snow strata in them. Stratigraphic units can be correlated from the veneer on the tundra to snow in the drift traps. The chief difference is that individual wind slabs may be 10 times thicker in the drift trap. Also, due to the great thickness of drifts (sometimes in excess of 5m), temperature gradients across drifts are considerably smaller, hence depth hoar development is reduced. Depth hoar strata found on the tundra are often absent in the drift sequence.



Weather Service records have underestimated snow precipitation on the Arctic Slope because of problems inherent in the measurement technique as pointed out by Black (1954). One of the reasons for the erroneously low values of winter precipitation on the Arctic Slope is due to the use of unshielded rain gages to measure the snow. This is interesting in light of research on shielding gages which has gone on for many years (Brooks, 1938; Warnick 1953). Also the low rate of snowfall combined with winds cause many entries of "trace" in the record (Jackson, 1960). At Barrow and Barter Island on Alaska's Arctic Coast the winter precipitation may contain as much as 80% trace in the record. The problems of measuring winter snowfall and determining the extent of wind transport and deposition of this snow on the Arctic Slope were discussed by Benson (1982). Measurement of the flux of windblown snow, combined with use of the Wyoming gage (Rechard and Larson, 1971) are useful in our attempts to determine the actual amount of snow precipitation on the Arctic Slope.

A significant fraction (up to half) of the snow which is precipitated on the tundra is relocated by wind. Part of the relocated snow is lost in transport by sublimation and part is concentrated in drifts. The part lost by sublimation remains to be determined, but the part which is deposited in drifts has been measured in selected "drift traps" since 1962. These traps are natural features, such as river banks and bluffs oriented normal to the wind. In general, we have chosen traps of sufficient size so that they do not fill to completion; thus, the volume of snow measured in the trap at the end of Winter is a measure of the total flux of snow blown from a given wind direction. Two sets of drift traps have been measured intensively. One is on high (10-20m) banks of the meandering Meade River near Atqasuk about 95 km SSW of Barrow, Alaska. It traps snow from both the east and west winds. The other is in the lee of some north-facing bluffs in the northern foothills of the Brooks Range at Imnavait Creek, near Toolik Lake about 200 km south of Prudhoe Bay, Alaska. These bluffs trap snow from southerly winds.

The two sets of drift traps highlight the major wind regimes of the Arctic Slope of Alaska. Along the northern coast and over much of the Arctic Slope the prevailing wind is from the East, with storm winds from the West. Near the Brooks Range the prevailing winds are from the South and are the result of katabatic drainage out of the mountains. We have attempted to define the northern boundary of wind transport from the south. Approximate values for the flux of wind-blown snow during a winter, stated in units of tonnes per meter, normal to the wind, is 72 from the east winds, 31 from the west winds (Bowling and Benson, 1982) and 20 from the south winds (Benson and Sturm, 1992). There is a wide variation in each of these values. Attempts to estimate wind transport of snow from wind speed records are underway (Wendler, 1978; Tabler and others, 1990).

The research at Imnavait Creek, mentioned above, is an interdisciplinary, ecological study funded by the U. S. Department of Energy. It is known as the R4D Research Project with its primary objective to determine the response of tundra to disturbance. It includes measurements of the amount and distribution of snow in the research watershed, as well as the structure of the snow, its biological role, its evolution during the melt season and the physical processes which govern snow melting (Kane and others, 1991). Snow maps with 5 cm contour intervals are being prepared for six consecutive winters, beginning with the 1984-85 winter. An example of these maps was presented by Liston (1986).

Snowmelt on the Arctic Slope is rapid; once it begins it is complete in a week or 10 days. However, the dates of initiation can vary by about a month from year-to-year. The primary control on spring warming is the progressive increase in solar radiation. Since this does not vary from one year to the next, the inter-annual differences in snowmelt must depend on secondary factors. These include the depth of the snowpack and the characteristics of advected air. Dr. Sue Ann Bowling of the Geophysical Institute found a

strong relationship between the start of melting and the thickness of the 1000-500 mb layer. This atmospheric thickness is directly proportional to the average temperature of the lower troposphere. It appears that the main mechanism of heat transfer to the snow is by long-wave radiation from the lower atmosphere. This also appears to be the primary mechanism of transferring heat from bare patches of the tundra to the surrounding snow covered areas. We have measured surface temperatures up to 30°C in bare patches of tundra, even though they are surrounded by snow with surface temperatures at 0°C. A more detailed study of these processes is underway (Bowling and Benson, 1992).

A study of snow chemistry was done along a transect from the Yukon River to Prudhoe Bay to describe the areal distribution of major ions and trace metals, and to discern the sources of impurities in the snow. A distinct break between cleaner snow to the south and dirtier, more variable snow chemistry to the north was found on the north side of the Brooks Range (Zukowski, 1989; Jaffe and Zukowski, 1992).

### **Passive Microwaves**

A study of passive microwave data derived from aircraft and satellites is underway by the author and colleagues at NASA's Goddard Space Flight Center and CRREL (Hall and others, 1989, 1991). We are examining an anomalous low-temperature region immediately north of the Brooks Range. Brightness temperatures are being plotted along selected traverses extending across Alaska from the Pacific Ocean to the Arctic Ocean. One data set spans the time in 1989 when all-time low January temperatures were followed within five days by all-time February high temperatures.

The role of depth-hoar in controlling microwave brightness temperatures has attracted attention for more than a decade (Hall and others, 1986). A detailed study of the specific effect of depth-hoar on passive microwave emission was done by removing sequential layers of the snowpack in interior and arctic Alaska. A maximum reduction in the effective emissivity was achieved by a depth-hoar layer 30 cm thick; layers thicker than this did not reduce the emissivity any further (Sturm, and others, 1992).

### **Perennial Dry Snow**

In the perennial snow of Alaska, there are regions of the dry snow facies (Benson, 1962; Williams and others, 1991) at high altitude in the Wrangell-St. Elias Mountains and in the Alaska Range. Our primary study of the dry snow facies in Alaska has been at the summit of Mt. Wrangell, where we are investigating the stratigraphy of dry snow and have begun some snow and ice coring. The stable isotope signal is not as straightforward as on the Greenland ice sheet. The reasons for this are being investigated. We are also using the melting of snow and ice on volcanoes as a calorimeter to study changes in volcanic heat flux. For example, the heat flux at the summit of Mt. Wrangell increased by two orders of magnitude after the 1964 earthquake and has melted over  $10^7$  m<sup>3</sup> of ice and snow from a single crater on the rim of the Summit Caldera (Wharton, 1967; Benson, 1968; Bingham, 1968; Motyka, 1983; Benson and others, 1985; Benson and Follett, 1986).

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# Estimation of Climatological Shortwave Albedo from Microwave Scattering Cross Sections at 5.3 and 9.25 GHz.

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## Introduction

In this research a series of electromagnetic interaction models are used to elaborate upon the role which geophysical properties play in defining microwave interaction with the snow covered sea ice. The primary focus here is on estimation of climatological shortwave albedo from the unique interaction of microwavelengths with the seasonally variable snow covered sea ice volume. Based on the experience gained from SIMS'90 and SIMS'91, the sensitivity analyses conducted here are considered typical of the SAR scattering seasons: Winter; Early Melt; Melt Onset; and Advanced Melt.

The scattering of microwave energy over sea ice is a complex function of the dielectric properties, surface roughness, and volume inhomogeneities of the snow and sea ice. The relative scattering cross section ( $\sigma^{\circ}$ ) changes over time and space. Spatial variability is largely a function of the geophysical properties which contribute to the volume dielectrics, or surface roughness of the material. Temporal variability is strongly controlled by the dielectric mismatch across the air-snow and snow-ice interfaces. The principal variables are the relative phases of water (ice, liquid and vapour), crystal size, brine volume and the vertical and horizontal roughness point estimates of the surface autocorrelation function.

Active microwave remote sensing consists of a sensor which generates and transmits (hence the term active) microwavelength energy towards a scattering surface over a range of incidence angles. This energy is scattered from the volume inhomogeneities and surface roughness characteristics of the earth material and is scattered. The synthetic aperture radar (SAR) is the most widely used form of active microwave remote sensing. Common frequencies for orbital and aerial SAR occur at the 5.3 GHz (C band) and 9.25 GHz (X band) frequencies. Polarizations on orbital, single frequency SAR sensors are typically HH or VV. This means that the signal is both transmitted and received with either a horizontal (HH) or vertical (VV) polarization.

The scattering mechanics of a SAR is a function of the sensor configuration, sensor-earth geometry, and dielectric properties of the material. Scattering can be separated into surface and volume components. If there is a strong dielectric mismatch at a particular interface then the surface scattering will dominate. The relative backscattering coefficient  $\sigma^{\circ}$ , is a measure of the amount of returned power per unit area, measured at the SAR antenna.

The relative complex dielectric constant [1] is used to express the permittivity ( $\epsilon'$ ) and loss ( $\epsilon''$ ) of the material relative to the frequency, polarization and incidence angle of the microwave energy.

$$\epsilon^* = \epsilon' + j\epsilon'' \quad [1]$$

Roughness can be approximated by the Fraunhofer Criterion [2]. This provides an index of roughness for a given vertical and horizontal dimension as a function of the angle and frequency of the incident microwave energy.



$$\sigma_h < \frac{\lambda}{32 \cos \theta} \quad [2]$$

A powerful tool for understanding the complexities of the microwave scattering process, and thereby constrain SAR algorithm development, is through the use of first order microwave scattering models. The term 'first order' is used because the models are able to handle only bulk volume attributes. The physics of the interactions are usually handled through radiative transfer (Mie or Rayleigh scattering), surface scattering theories (of the Kirchhoff type) and dielectric mixture models, each of which require geophysical variables as inputs.

This is a forward modelling approach, where the physical properties of the snow and sea ice are known, and the electromagnetic (EM) interaction is predicted. The use of forward modelling allows for an improved understanding of the scattering process through an analysis of individual geophysical variable contributions to the magnitude and seasonal evolution of  $\sigma^\circ$ . Modelling also allows for interpolation of  $\sigma^\circ$  within a constrained range of geophysical observations acquired over a space/time continuum. The final attribute of the modelling process is the capability to extrapolate beyond the spatial and temporal sampling boundaries inherent in the surface data collection.

## Methods

The research question of interest here is addressed by linking  $\sigma^\circ$  to measured daily averages of climatological shortwave radiation over a seasonally evolving snow covered sea ice volume. Quadratic models are computed to link averaged  $\sigma^\circ$ , modeled from the temporal evolution of geophysical properties, to the daily averaged integrated climatological albedo ( $\alpha$ ). Data from the SIMS'91 automated station are used to compute  $\alpha$ . Physical properties, obtained from the snow pit sampling on all first year ice sites, for Julian Days 138 to 163, provided the inputs for the microwave scattering models. Specific details of the methods for measuring physical properties can be found elsewhere (Barber et al. 1992)<sup>1</sup>.

Model trials were computed at 5.3 and 9.25 GHz, at incidence angle of 20°, 30° and 40°, at HH polarization. A Least Squares Polynomial regression was used to generate the quadratic models. Appropriateness of the models were assessed using the Durban-Watson metric, and by visually examining a plot of the residuals versus predicted values. Statistical diagnostics are provided for each model computed. The Least Squares Polynomial prediction intervals for  $\alpha$ , based on a 90 percent confidence level provide a specific indication of the precision associated with prediction of  $\alpha$  from an existing observation of  $\sigma^\circ$ .

## Results

Results indicate (Figure 1) that a strong statistical relationship existed between the observed daily averaged albedo ( $\alpha$ ) and the modeled relative scattering cross section ( $\sigma^\circ$ ). Since  $\sigma^\circ$  was modeled from daily averages of measured geophysical properties, the observed relationship is considered valid over the spatial and temporal scales typified by the SIMS'91 study site.

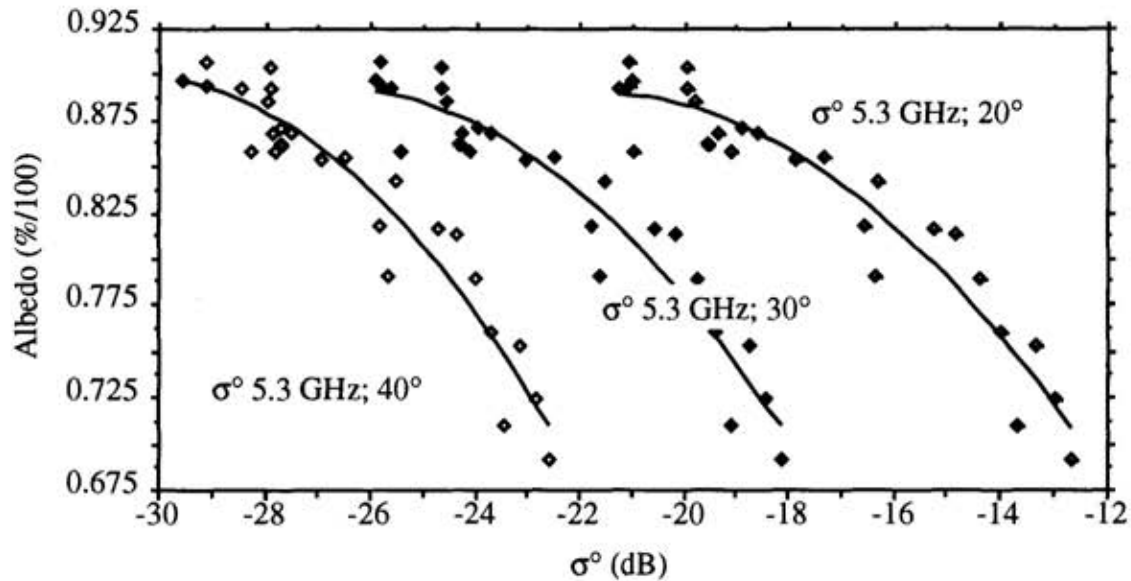
Results from the 5.3 GHz frequencies (Figure 1) indicate that  $\sigma^\circ$  increased as integrated climatological shortwave albedo ( $\alpha$ ) decreased. For a change in  $\alpha$  of 0.22 there was a corresponding increase in  $\sigma^\circ$  from -21 dB to -12 dB, at a 20° incidence angle.

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<sup>1</sup> Barber, D.G., D.G. Flett, R.A. De Abreu, and E.F. LeDrew. 1992. Spatial and Temporal Variations in Sea Ice Geophysical Properties and Microwave Remote Sensing Observations: The SIMS'90 experiment. *Arctic*. In Press (Sept. issue).



Note that the daily averaged  $\alpha$  were obtained from a snow cover throughout the duration of SIMS'91 (i.e., the snow pack never metamorphosed completely to a melt pond). The primary physical variables causing the increase in  $\sigma^\circ$  were the increased  $W_v$  of the snow pack and a larger contribution to  $\sigma^\circ_v$  from the increase in the snow crystal radius.



**Figure 1.** Least squares polynomial regression model fits for daily averages of the relative scattering cross section ( $\sigma^\circ$ ) versus daily averages of the integrated climatological albedo ( $\alpha$ ). Scattering model results are computed for 5.3 GHz, HH polarization, at 20°, 30° and 40° incidence angles.

Statistical diagnostics (Table 1) and visual examination of the scatter plots suggest that a least squares polynomial regression model would be appropriate for these data. Results from the Durban Watson test ( $D_w$ ) and examination of residual plots indicated no apparent problems in the assumption of independence of the error terms, required as a validity condition for these quadratic models. A test of the curvature of the least squares fit ( $\sigma^{\circ 2}=0$ ) indicated that for each model the  $\sigma^{\circ 2}$  term was appropriate. For each of the incidence angles a strong coefficient of determination ( $R^2$ ) was observed. The resulting models can be used to predict albedo based on the observed  $\sigma^\circ$  at the prescribed frequency, polarization and incidence angle.

**Table 1.** Statistical diagnostics from the least squares polynomial regression for  $\alpha$  on  $\sigma^\circ$  at 20°, 30° and 40° incidence angles at 5.3 GHz, HH polarization.

Frequency	Incidence	$R^2$	$D_w$	$\sigma^{\circ 2}=0$	F-Stat
5.3 GHz	◆ 20°	0.92	1.716*	no	131.86
5.3 GHz	◆ 30°	0.92	1.718*	no	131.64
5.3 GHz	◆ 40°	0.918	1.684*	no	129.53

An intercomparison of the three 5.3 GHz models (Table 2) indicated that the slope of the models showed an average increase (i.e., increased negative slope from -0.095 to -0.182) with an increase in the angle of incidence. The curvature of the models showed little change over the three incidence angles. The larger slope coefficient at the 40° incidence angles would increase the separability of  $\sigma^\circ$  and  $\alpha$ . There is a tradeoff however, in that the larger slope is also coupled with a smaller intercept, which means the average  $\sigma^\circ$  is about 10 dB lower than the

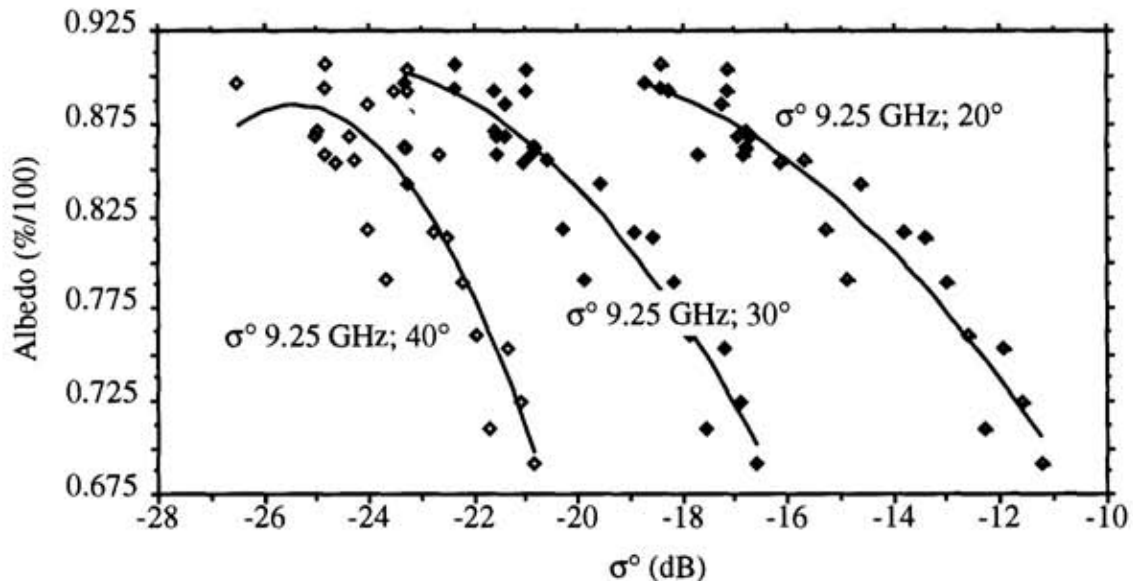
equivalent 20° incidence angle model. This is significant in microwave sensors with a high noise floor, such as ERS-1.

At 5.3 GHz, prediction of  $\alpha$  from  $\sigma^\circ$  would be equally precise at 20°, 30° and 40° incidence angles. Based on these data a minimum average prediction interval ( $CI_{\min}$ ) for an estimated  $\alpha$ , would range from 7.8 to 7.9 percent (Table 2). The maximum prediction interval ( $CI_{\max}$ ) ranged from 8.7 to 8.8 percent. These intervals were computed for a 95 percent confidence level against a Type I error and the variance term, used in computation of the confidence interval, was computed for an existing rather than new observation of  $\sigma^\circ$ .

**Table 2.** Mean squared error, minimum and maximum confidence intervals and statistical models for the prediction of  $\alpha$  from  $\sigma^\circ$ , at 20°, 30° and 40° incidence angles at 5.3 GHz.

Frequency	MSE	$CI_{\min} \hat{\alpha}_{0.95}$	$CI_{\max} \hat{\alpha}_{0.95}$	Model
5.3 GHz	0.0003429	0.078	0.087	$\hat{\alpha} = -0.141 - 0.095\sigma^\circ - 0.002\sigma^{\circ 2}$
5.3 GHz	0.0003434	0.079	0.087	$\hat{\alpha} = -0.853 - 0.13\sigma^\circ - 0.002\sigma^{\circ 2}$
5.3 GHz	0.0003486	0.079	0.088	$\hat{\alpha} = -1.878 - 0.182\sigma^\circ - 0.003\sigma^{\circ 2}$

Results from the 9.25 GHz frequencies (Figure 2) indicate that  $\sigma^\circ$  increased as integrated climatological shortwave albedo decreased. For a change in  $\alpha$  of 0.22 there was a corresponding increase in  $\sigma^\circ$  from -18 dB to -11 dB, at a 20° incidence angle. The magnitude of  $\sigma^\circ$  at 9.25 GHz was larger than the corresponding 5.3 GHz frequencies because of the increased magnitude of  $\sigma_s^\circ$ , from the snow surface, to  $\sigma^\circ$ . The slopes of the models were smaller in the 9.25 GHz data, primarily because of the compensatory effect of  $\sigma_s^\circ$  and  $\sigma_v^\circ$  on  $\sigma^\circ$ .



**Figure 2.** Least squares polynomial regression models for daily averages of the relative scattering cross section ( $\sigma^\circ$ ) versus daily averages of the integrated climatological albedo ( $\alpha$ ). Scattering model results are computed for 9.25 GHz at 20°, 30° and 40° incidence angles.

Statistical diagnostics (Table 3) and visual examination of the scatter plots suggest that the 20° and 30° incidence angles, at 9.25 GHz, were appropriately modeled with a quadratic function and that the 40° incidence angle was not ( $D_w=0.891$ ). Tests for the significance of the  $\sigma^2$  term showed that the curvature of the lines were each significantly different from zero. Although not as strong as the 5.3 GHz data, the 9.25 GHz results indicated good correlations between  $\sigma^\circ$  and  $\alpha$ . The coefficient of determination ranged from 0.919 to 0.762.

**Table 3.** Statistical diagnostics from the least squares polynomial regression for  $\alpha$  on  $\sigma^\circ$  at 20°, 30° and 40° incidence angles at 9.25 GHz.

Frequency	Incidence	R <sup>2</sup>	Dw	s <sup>o2=0</sup>	F-Stat
9.25 GHz	◆ 20°	0.919	1.824*	no	129.91
9.25 GHz	◆ 30°	0.901	1.584*	no	104.20
9.25 GHz	◆ 40°	0.762	0.891	no	36.89

An intercomparison of the three 9.25 GHz models (Table 4) indicated that the slope of the models showed an average increase (i.e., increased negative slope from -0.091 to -0.458, with an increase in the angle of incidence). The curvature of the models showed a small change between 20° and 30° and a larger change between 30° and 40° incidence angles.

At 9.25 GHz, prediction of  $\alpha$  from  $\sigma^\circ$  would be equally precise at 20° and 30° incidence angles. The 40° incidence angle at 9.25 GHz appeared to provide significantly less predictive capabilities than either the 20° and 30° incidence angles at 9.25 GHz or all incidence angles at the 5.3 GHz frequencies. Based on these data a minimum average prediction interval ( $CI_{min}$ ) for an estimated  $\alpha$ , would range from 7.9 to 13 percent (Table 4). The maximum prediction interval ( $CI_{max}$ ) ranged from 8.9 to 17 percent.

**Table 4.** Mean squared error, minimum and maximum confidence intervals and statistical models for the prediction of  $\alpha$  from  $\sigma^\circ$  at 20°, 30° and 40° incidence angles at 9.25 GHz and an HH polarization.

Frequency	MSE	$CI_{min} \hat{\alpha}_{\sigma^{ss}}$	$CI_{max} \hat{\alpha}_{\sigma^{ss}}$	Model
9.25 GHz	◆ 0.0003476	0.079	0.089	$\hat{\alpha} = -0.036 - 0.091\sigma^\circ - 0.002\sigma^{\circ 2}$
9.25 GHz	◆ 0.0004249	0.088	0.10	$\hat{\alpha} = -1.009 - 0.155\sigma^\circ - 0.003\sigma^{\circ 2}$
9.25 GHz	◆ 0.001	0.13	0.17	$\hat{\alpha} = -4.925 - 0.458\sigma^\circ - 0.009\sigma^{\circ 2}$

## Conclusions

Results indicate that both 5.3 and 9.25 GHz frequencies, at HH polarization and incidence angles of 20°, 30° and 40° could be used to estimate the daily averaged integrated climatological albedo ( $\alpha$ ). Least Squares Polynomial regression models were constructed for combinations of frequency and incidence angle. The models at 5.3 GHz, HH polarization, at 20°, 30°, and 40° incidence angles were equally precise in predications of  $\alpha$ . The coefficient of determinations for these three models were 0.92, 0.92 and 0.918. The models at 9.25 GHz were slightly less precise, particularly at the 40° incidence angle. The coefficient of determination for the 20°, 30° and 40° incidence angles were 0.92, 0.90, and 0.762. The reduction in precision at the 40° incidence angles was attributed to the increased sensitivity at both 5.3 and 9.25 GHz, to the snow surface scattering term ( $\sigma^{ss}$ ) used in computation of the total relative scattering cross section ( $\sigma^\circ$ ).

The relevance of the  $\alpha$ - $\sigma^{\circ}$  model is that the daily averaged albedo could be computed from daily observations with orbital SAR sensors, independent of the ubiquitous cloud cover, typical of this component of the Earth-Atmosphere system. Since the radiation balance is the major contributor to the total energy balance, this result is considered as a significant step towards remote measurement of components of the total sea ice energy balance.

# Creating Temporally Complete Snow Cover Records Using A New Method for Modelling Snow Depth Changes

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## Abstract

The availability of a new daily climate data set provides a unique opportunity to use daily snow cover data from multiple weather stations for regional climate analysis. Here, this set is used in the development of an empirically-based statistical method for modelling changes in snow depth for the Great Plains region of the United States. Long-term daily data from 62 cooperative weather stations with records extending back to the turn of the century are used. This method offers an alternative to traditional energy balance or temperature index methods that are often inaccurate for regions prone to shallow and ephemeral snow cover. Average daily change in snow depth at a given mean air temperature is computed for Winter and Spring for four states in the central and northern Great Plains. These changes in depth with temperature show considerable variation, both regionally and seasonally, increasing from the northern plains to the central plains, and from Winter to Spring. The relationship between snow depth change and mean temperature may be described statistically by a second order polynomial. The derived equations are used in conjunction with temperature and snowfall data to simulate snow pack behavior at four control stations in the region. Results from a sensitivity analysis indicate that gaps of varying lengths may be filled with a high degree of confidence to create temporally complete snow cover data at a station. Other uses for this methodology include simulation of snowpack behavior in regional climate change studies, and estimating basin runoff for hydrologic and agricultural studies.

## Introduction

In the northern and central Great Plains of the United States and the adjacent Canadian prairie, snow represents an important hydrologic resource. It is estimated that snowfall in this region, due in part to its lower evaporation potential, is twice as valuable as a resource for commercial agriculture as rain (Greb, 1980). For example, over the Canadian prairie, snow represents only 30% of the annual precipitation, but accounts for greater than 80% of the surface runoff (Gray & Landine, 1988). Along with the hydrologic implications of snowfall in this region, climate change scenarios show an amplification of anthropogenically-induced warming in the plains region where snow cover is ephemeral (Manabe & Wetherald, 1980; Hansen, 1986). The ability to understand and model changes in the depth of prairie snowpacks is therefore important for adequate resource management, as well as for regional climate change analysis.

Two methods commonly used for estimating snow depth and snowmelt are energy balance and temperature index approaches. Energy balance models attempt to solve the fundamental equations of energy, mass and momentum. Solving these equations involves input such as solar radiation, wind speed, water equivalent of snow and net short and longwave radiation components. Much of this information is unavailable or difficult to measure for large scale operational use (Gray & Male, 1981). The temperature index



method is based on a simple relationship between snowmelt and temperature above a given base temperature (usually 0°C). A melt factor term is defined to account for such things as site exposure, season, ground cover, cloudiness and the albedo of the snow. Problems with this method occur from assumptions made when defining regionally specific melt factors and an appropriate base temperature for the initiation of melt (Winston, 1965; Huber, 1983; Kuhn, 1987; Kondo & Yamazaki, 1990). In addition, the energy balance and temperature index methods are more accurate in forested alpine environments where snow cover is deep and present throughout the snow season (Granger & Male, 1979; Gray & Landine 1988). The application of both modelling approaches to the shallow snows in the prairie environment often proves to be inaccurate (Gray & Landine, 1988).

The development of our Depth Change (DC) method offers a two-step approach for modelling snowpack behavior in the Great Plains. The first step is the computation of average daily changes in snow depth for a given mean temperature using long-term historical snow cover data from multiple weather stations in a region. The availability of hundreds of cases of snow depth change from a dense network of climate stations permits the development of statistically reliable regression equations. Step two uses these equations in conjunction with snowfall data, as a means for both accumulating and decreasing the amount of snow cover on the ground. This permits the filling of missing daily snow cover data at individual climate stations, and may be used as a means for simulating regional snowpack behavior for use in climate change studies. Other uses include estimating water yields for hydroelectric power generation, forecasting streamflow volume and estimating soil moisture (Granger & Male, 1979; Gray & Landine, 1988; Carr, 1989).

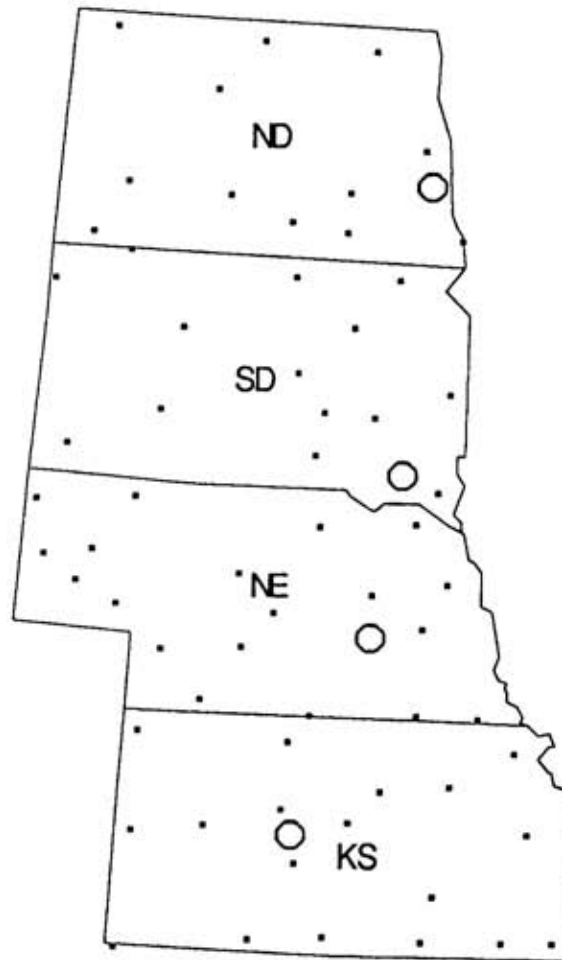
### **Daily Historical Climate Data Set**

The use of snow cover in climate change research has until recently been limited by the lack of long-term snow cover data of a high quality (Karl et al., 1989). The Daily Historical Climate Data Set compiled by Robinson (1988) provides the first daily accounting of snow cover for approximately 1100 cooperative weather stations distributed throughout the United States. Along with snow cover, each record provides daily data on snowfall, precipitation and maximum and minimum temperatures. The digital data are carefully quality controlled. They provide a unique opportunity for studies regarding long-term temporal and spatial variability of snow cover over the Great Plains.

Approximately 100 stations in the set are located in North Dakota, South Dakota, Nebraska, and Kansas. Records at these sites extend back to the turn of the century, however, even at the best climate stations between 2% and 8% of the snow cover data are missing. The amount of missing snow cover data from October-April for all stations in the region ranged from a low of 0.5% to a high of 57%, with 76% of the stations having less than 15% missing. Gaps in snow cover in a preliminary analysis of snow conditions in this region were simply filled by a nearest neighbor substitution method on a seasonal basis (Robinson & Hughes, 1991). However, given the spatial variability of snow cover in this region, this procedure is at times unreliable.

### **Depth Change (DC) Method**

The central and northern Plains states of North Dakota, South Dakota, Nebraska and Kansas are considered subregions in this study. Each is approximately 3° in latitude by 7° in longitude. From 12 to 19 well-distributed stations in each state are used for the development of regression equations that are based on a comparison of daily changes in snow depth and mean temperature (fig. 1). One station in each region, independent of the



**Figure 1.** Distribution of Great Plains climate stations employed in this study. Stations marked with opened circles are study control stations.

original group, is used as a control station to evaluate the efficacy of using the equations to estimate changes in snow depth given a mean temperature. Snow depths between 2.5 and 65 cm are considered in this analysis. Unlike the temperature index method, our routine does not arbitrarily take 0°C as the base temperature for the initiation of snowmelt. Rather it is observed that above a mean temperature of -10°C changes in depth/day are of some importance in dictating the depth of a snowpack when observed over a period of days or weeks. The snow season is divided into two portions, the first from November through mid-February, and the second from mid-February through April.

The first step of the DC method is the determination of the change values, and associated regression equations. To accomplish this, a computer program (CHANGEVALUE) computes the average daily change in snow depth for a given mean temperature. The program allows the user to define a particular region and season to examine. For each station in a given region, the previous day's snow depth is compared to the current day's snow depth. If a decrease in depth has occurred, the change is stored along with the mean daily temperature. Days with observed snowfall are omitted from the analysis. When all the stations within the region are scanned, the average change in snow depth/mean temperature (given that a change has occurred) is computed. The values are then adjusted based on the number of times snow cover is observed and no change in snow depth occurs at that temperature value. The average number of cases found for each 0.56°C (1°F) temperature increment is approximately 180 for the Winter snow season, and 140 for the Spring. A decrease in the number of cases occurs at both ends of the temperature scale, most notably in the winter at high temperatures (Table 1). Thus, the derived change values are considered suspect at high temperatures. A regression analysis follows, using the change values and associated mean air temperatures to develop statistical snow depth change models. Eight models have been developed for this study representing the two seasons for each of the four states.

The second step of this method involves using the regression equations to estimate snow cover depths at a station. A program called DATAFILL reads every record for a given climate station and searches for missing snow cover data. If a missing value is found the program attempts to estimate it in one of three ways. If the only possible snow cover value is zero based on temperature, snowfall, and snow cover from the previous and current day's record, this is filled in first. If the mean temperature for the day is below -10°C, the current day's snow depth is set equal to the previous day's snow depth. If the mean temperature is greater than -10°C, the current snow depth value is estimated to be the previous day's snow depth plus the current day's snowfall, less the change in snow depth determined by solving the regression equation at that temperature. In this way, both daily accumulation and decreases in snow cover are estimated.

**Table 1.** Sample of number of cases/mean temperature (°C) found for Winter (W) and Spring (S) snow seasons.

Temperature	ND (W)	ND (S)	SD (W)	SD (S)
-11.1	42	31	51	39
-10.0	47	39	75	69
-8.9	60	55	104	97
-7.8	81	72	122	119
-6.7	108	97	171	159
-5.6	130	148	213	171
-4.4	165	155	275	169
-3.3	154	184	304	232
-2.2	163	224	285	245
-1.1	173	212	298	277
0	160	243	249	264
1.1	111	216	215	254
2.2	60	166	146	200
3.3	41	132	93	154
4.4	29	98	48	103
5.6	4	75	34	77

Temperature	NE (W)	NE (S)	KS (W)	KS (S)
-11.1	105	31	71	23
-10.0	130	51	71	17
-8.9	180	74	112	29
-7.8	205	96	147	37
-6.7	235	130	160	55
-5.6	296	148	191	77
-4.4	336	170	215	68
-3.3	349	190	269	79
-2.2	374	204	277	111
-1.1	365	240	276	114
0	350	231	238	124
1.1	294	240	236	149
2.2	258	204	198	140
3.3	160	156	165	99
4.4	97	116	92	74
5.6	51	97	72	59

Two related issues have been addressed in the development of this method, the first being the settling of a fresh snowpack, the second being the affect of rain on snow depth. The influence of settling on snow depth is examined by running the DC method for depth classes of 2.5-25 cm and 25-65 cm at North Dakota stations. While rapid settling is found to occur within the first few days following a snowfall, there are too few cases of snow depth change at most temperatures to quantify this impact. Similarly, although rain can have a significant impact on the depth of snow, too few cases are available in the Great Plains to adequately quantify this impact. The occurrence of these events is not excluded from the determination of the change values derived in this study.

### Regional And Seasonal Variations In Change Values

For Winter and Spring in the Great Plains, the decrease in snow depth at a given mean temperature gets progressively larger as one continues southward. For example, at -4°C in Winter, depth decreases by 0.89 cm/day. The values for South Dakota, Nebraska, and Kansas are 1.24 cm/day, 1.71 cm/day, and 1.75 cm/day respectively. Also, winter values are approximately 1.5 times lower than spring values (figs. 2 & 3). For example, at -8°C in North Dakota, depth decreases by 0.36 cm/day in Winter and 0.8 cm/day in Spring. At 4°C the decrease is 2.89 cm/day and 4.26 cm/day respectively. Similar results were found for Kansas, where depth decreases at -8°C vary from 1.20 cm/day in Winter to 2.06 cm/day in Spring. Associated values at 4°C are 3.32 cm/day and 4.89 cm/day respectively (Table 2).

The relationship between snow depth change and mean temperature for all cases is best described by a second order polynomial (fig. 4).  $R^2$  values range from a low of 0.98 in North Dakota in Winter to a high of 0.99 in Kansas in Winter, and all are significant at the 99% level. The statistical models describe the small changes in snow depth that occur at mean temperatures far below freezing, encompassing the phenomena of melt, compaction and sublimation. As mean temperatures increase, so do the daily changes in snow depth. As mentioned previously, at mean temperatures above 3.5°C, the number of cases of decreasing snow depth diminishes, particularly in Winter. This causes greater variability in computed change values. At these temperatures some of the variability results from shallow snowpacks melting away completely before the full potential of the day's temperature to melt snow is reached. An observer bias is also found for some stations,

where a tendency to record 2.54 cm (1 inch) of snow on the ground existed despite persistent mean temperatures above 3.5°C.

**Table 2.** Sample of regionally and seasonally derived snow depth change values for -8°C, -4°C, 0°C, and 4°C

Temperature=-8°C

Region	Winter (cm/day)	Spring (cm/day)
North Dakota	0.36	0.80
South Dakota	0.57	1.14
Nebraska	1.10	1.91
Kansas	1.15	2.06

Temperature=-4°C

Region	Winter (cm/day)	Spring (cm/day)
North Dakota	0.89	1.66
South Dakota	1.24	2.05
Nebraska	1.71	2.79
Kansas	1.75	2.85

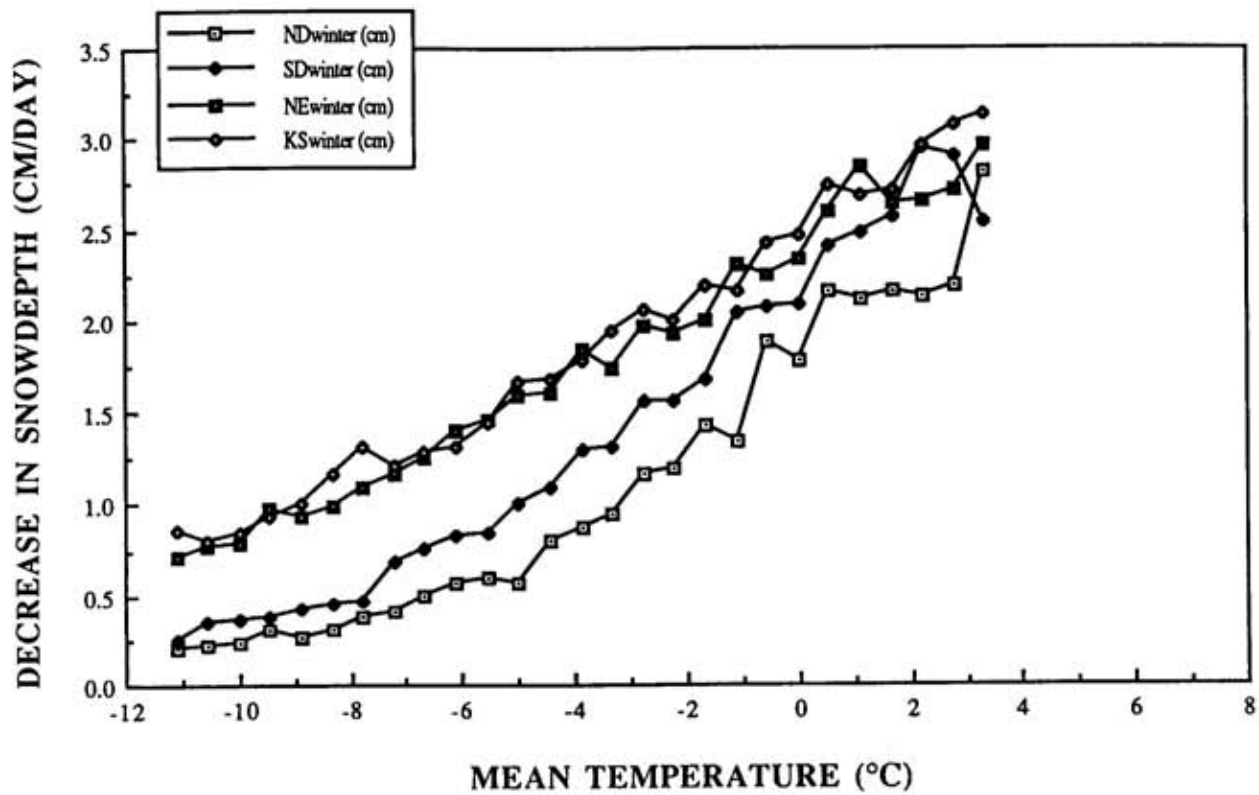
Temperature=0°C

Region	Winter (cm/day)	Spring (cm/day)
North Dakota	1.73	2.81
South Dakota	2.12	3.19
Nebraska	2.37	3.69
Kansas	3.08	4.62

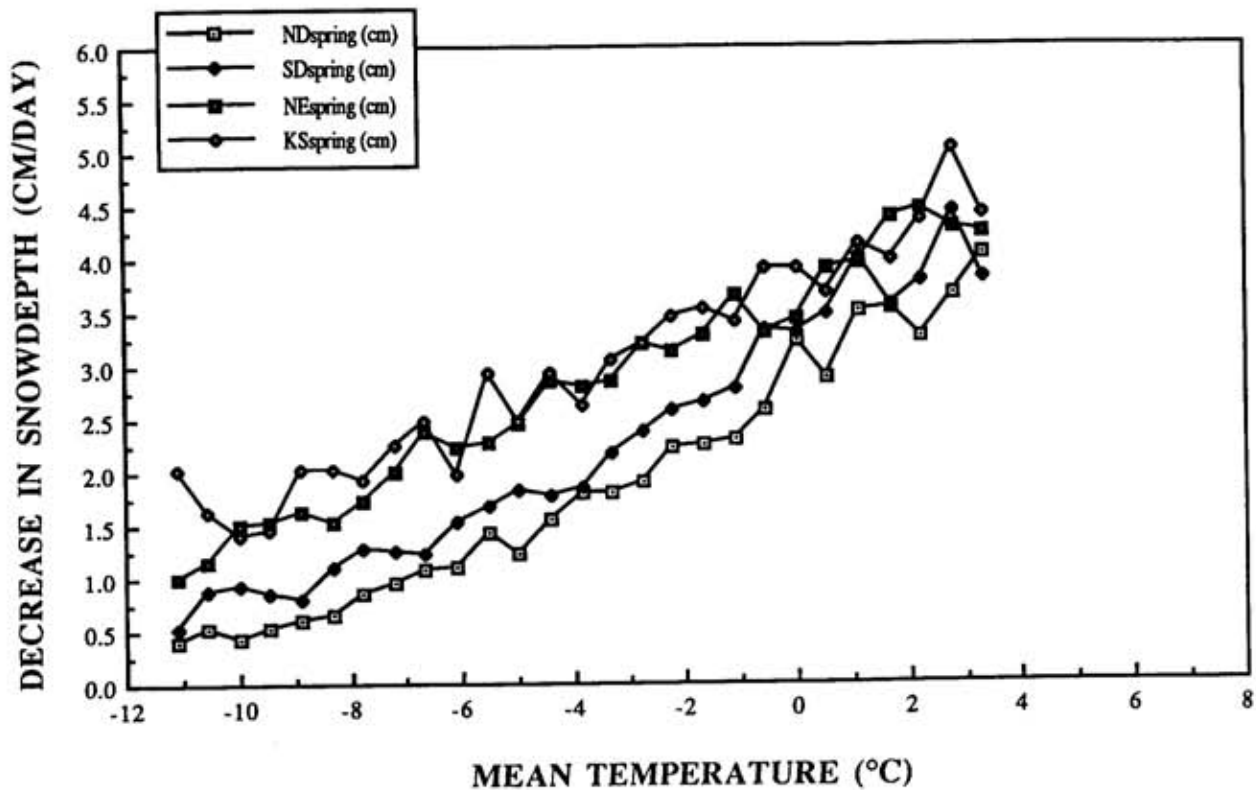
Temperature=4°C

Region	Winter (cm/day)	Spring (cm/day)
North Dakota	2.89	4.26
South Dakota	3.20	4.56
Nebraska	3.08	4.62
Kansas	3.32	4.89

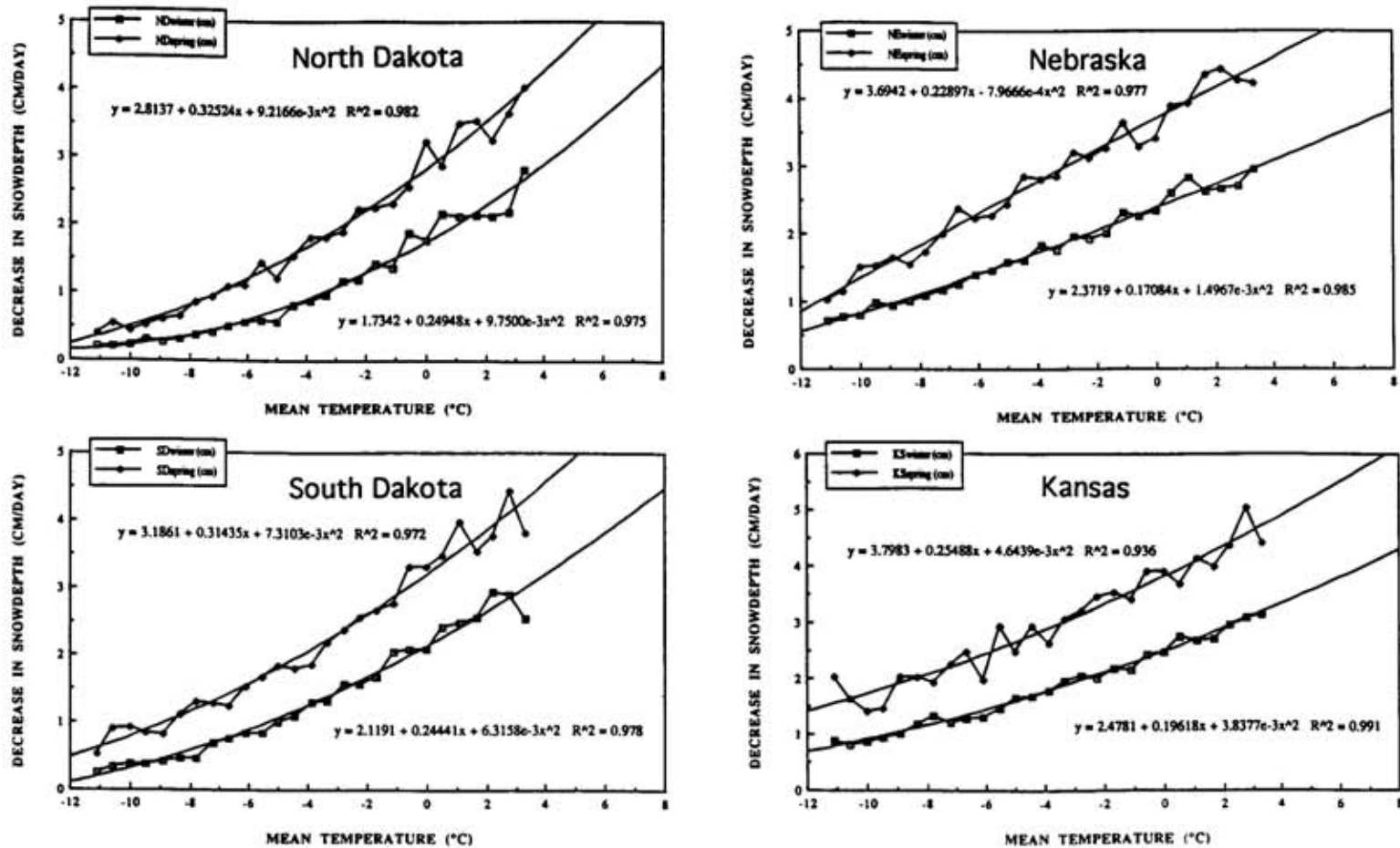




**Figure 2.** Association between daily snow depth change and mean temperature during Winter in the four study states.



**Figure 3.** Association between daily snow depth change and mean temperature during Spring in the four study states.



**Figure 4.** Association between daily snow depth change and mean temperature during Winter and Spring for the four study states. Observed results and a second order polynomial fit to these data are shown.

## Evaluation Of The Depth Change Method

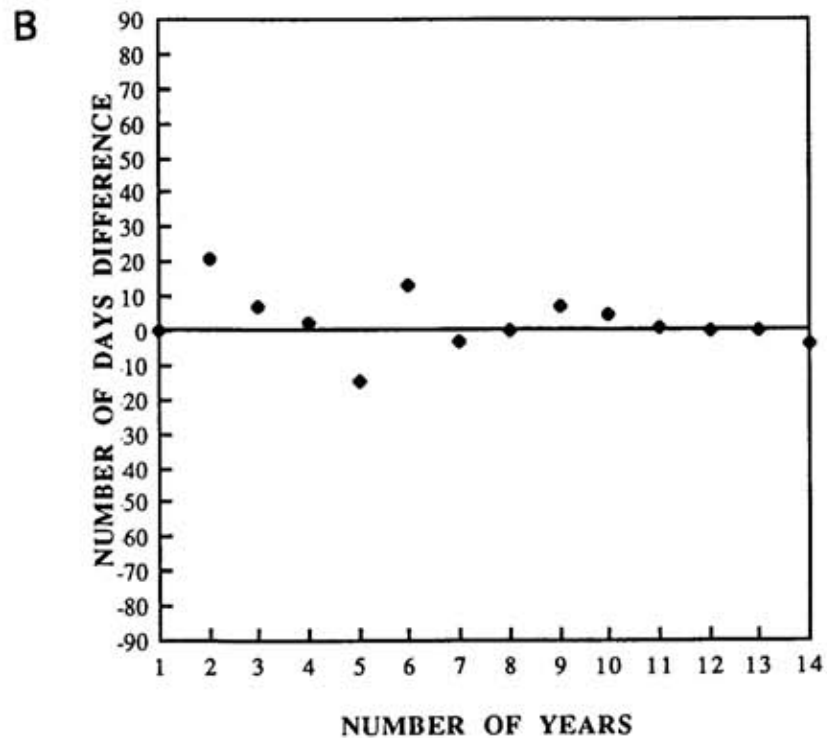
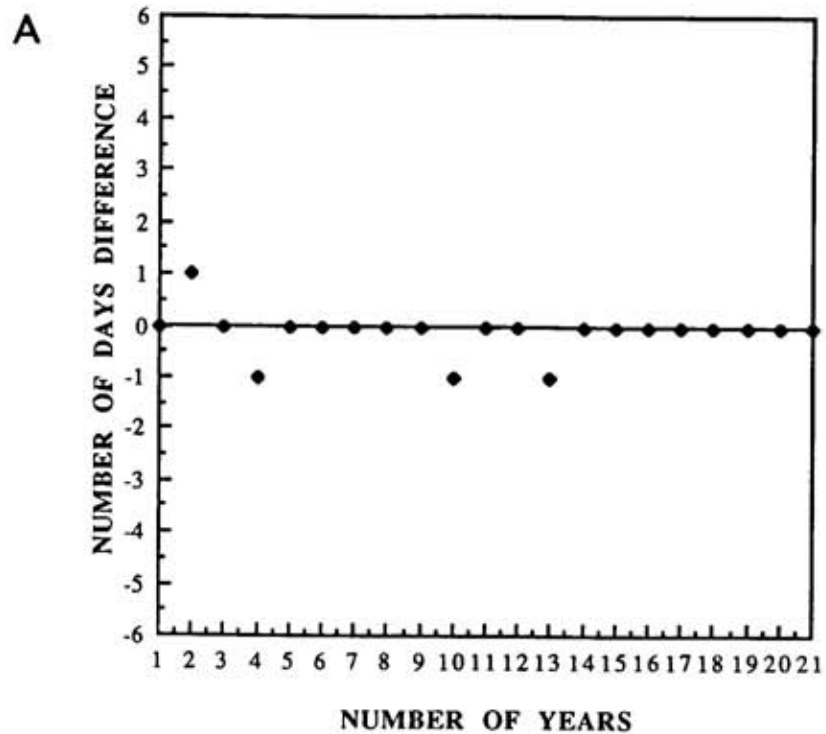
To evaluate the use of the DC method for data filling purposes, a sensitivity analysis is done using one station within each of the four states (fig. 1). These stations are not included in the formation of the regression model. A stepwise approach is used where initially two days in each month are changed from the observed snow depth value to a missing value. The missing days are then estimated by using observed temperature and snowfall data along with the appropriate regression model. Observed snowfall values are added to the existing snowpack, thereby allowing increases in the daily snow cover. Decreases in the snowpack occur through use of the temperature data and the regression equation. Once filled, the original file and filled data file are compared and differences between the two files quantified. This procedure has been repeated with 4, 8, 16, and 31 days/month changed to missing and then subsequently filled.

Results of this sensitivity analysis are illustrated using the control stations located in Amenia, ND, and Bison, KS. Comparisons between the filled file and the original file are done by summing the total number of days in Winter (DJF) with more than 2.5 cm (1 inch) of snow on the ground and differencing the numbers. The two cases shown for each station represent data filling with two days in every month filled and all days in every month filled, the latter being a complete simulation of snowpack behavior (figs. 5 & 6). Therefore, each point on the graph represents the difference in the total number of days with snow on the ground between the files for an entire Winter. The y-axis is scaled by the maximum number of filled data points in the season. For Amenia, ND with two days/month filled, there are 21 years of winter data compared, or a total of 126 days of estimated snow cover data. When the observed and modeled files are compared, only four days exist where one file indicates snow on the ground and the other does not (fig. 5a). In the case of simulating the entire winter snow season for Amenia, 14 years of snow cover data are compared to the original file. Even in this extreme case, the modeled snowpack is quite close to the actual observed conditions (fig. 5b). The largest difference between files in a year is 20 days. Of the 14 years compared, 11 years, representing 992 filled days, have fewer than 7 days/season difference. For Bison, KS the southernmost control station, 37 years of original data are compared to the filled snow cover data. For the case where two days/month are filled, only 7 of 222 modeled days indicate snow on the ground where the observed does not (fig. 6a). When the entire winter snow cover is simulated, only 3 of 37 years have fewer than one week difference between observed and modeled files (fig. 6b).

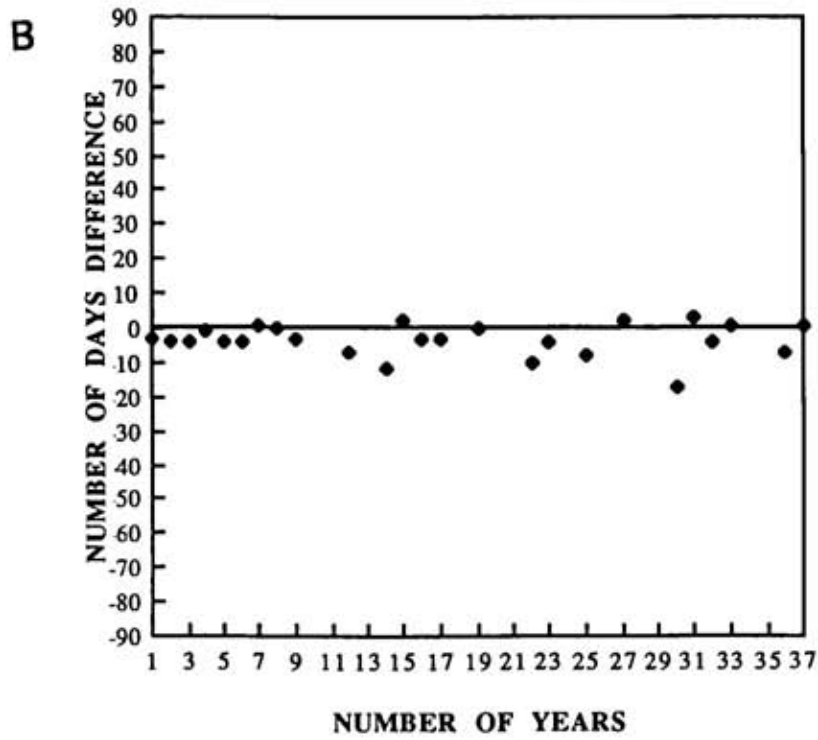
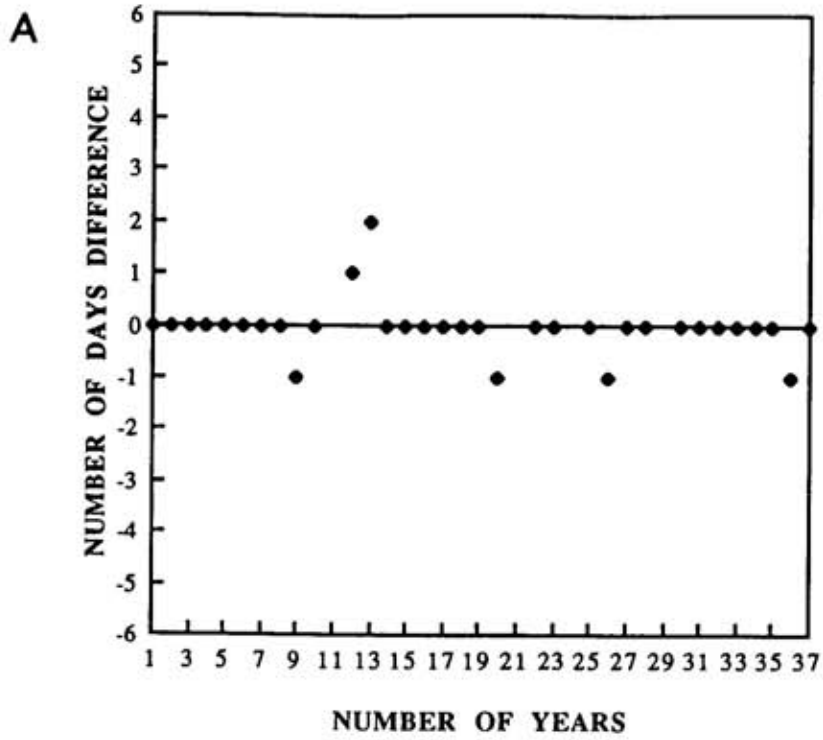
Analysis of the control stations located at Menno, South Dakota and Central City, Nebraska shows similar favorable results. For all cases, use of the regression model in conjunction with observed snowfall and temperature values proves to be an acceptable means for filling gaps of varying length within the data set.

## Discussion

The DC method approaches the problem of modelling snow depth change in a purely statistical manner. Unlike the energy balance and temperature index methods for modelling *snowmelt*, the DC approach models changes in *snow depth*. This does not fully explain the success of the DC method in modelling Great Plains snow cover where the others often fail. The melt and depth difference is only important at subfreezing temperatures, where depth changes slowly. Other reasons why traditional snowmelt models are less reliable in this region are illustrated from the results of the following studies. McKay (1964) used a degree-day approach with station and snow survey data to evaluate snowpack



**Figure 5.** Comparisons between observed and modeled (filled) values of the number of days with  $\geq 2.5\text{cm}$  of snow cover during the Winter (DJF) at Amenia, ND. Differences between the original and filled (filled - original) when two days per month are filled are shown in A. B shows the same, except all days are filled.



**Figure 6.** Same as fig. 5, except for Bison, KS.



characteristics in Saskatchewan and Manitoba, and found problems with appropriately defining the base air temperature for initiation of melt. Granger and Male (1979) used an energy budget approach to model snowmelt in the Bad Lake Watershed in Saskatchewan using temperature, humidity and vertical wind speed data for three years from 1974, 1975, and 1976. A significant problem arose due to the affect of advection of heat from the ground in areas of shallow and patchy snow. This has been noted as well by Heron and Woo (1978), Male (1980), Aguado (1983), and Gray and Landine (1988). Meier (1985) noted that modelling the seasonality of snowmelt behavior as well as being able to precisely measure all the energy fluxes involved with snowmelt is impossible for operational use at most locations.

Gray and Male (1981) cite studies by Yoshida (1962), Anderson (1973) and the Army Corps of Engineers (1956), where melt factors ranged from 0.39-0.80 cm/°C-day in the Tadami River basin in Japan to 0.13-0.37 cm/°C for sites in the United States. Thus, at 3°C these factors predict changes in snow depth ranging from 0.39 cm/day to 2.40 cm/day. This compares with depth changes in Winter at 3°C computed using the DC method ranging from 2.57 cm/day in North Dakota to 3.10 cm/day in Kansas. In Spring, these values range from 3.87 cm/day to 4.60 cm/day.

Since the DC method computes the differences in snow depth change values empirically, regional and seasonal variations in snowpack behavior are preserved. Differences found in both dimensions are most likely due largely to variations in solar radiation reaching the surface of the Earth. As one progresses northward, the shorter day length and higher solar zenith angles during the snow season result in smaller changes of snow depth at the same mean temperature. The greater difference in depth changes between South Dakota and Nebraska compared to between North and South Dakota and between Nebraska and Kansas may reflect differences in the persistence of winter snow cover in the northern and southern halves of the study area. Other reasons for regional and seasonal variations in change values may be related to differences in antecedent soil moisture, the number of full melt events that occur in a snow season and in meteorological conditions such as winds and clouds. Whether the soil underlying the snowpack is frozen or unfrozen may also in part account for variations.

## **Conclusions**

The Depth Change (DC) method is successful at modelling seasonal changes in snow cover across the Great Plains. This method is more reliable than the energy balance and the temperature index methods because it relies on daily historical snow cover data to build statistical snow depth change models. Change values, not constrained by predefined limits, are allowed to vary considerably over a given region, allowing a flexible approach to modelling snowpack conditions. In conjunction with snowfall data, one can quite accurately recreate or fill in missing snow cover values at individual stations on a daily basis. This is important as a means to permit detailed investigations of long-term variations or potential trends in snow cover over the Great Plains, a region models suggest will be significantly affected by climate changes associated with increased levels of greenhouse gases.

## **Acknowledgments**

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# The Need for Removing Biases from Rain and Snowgauge Measurements

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Over the past several years, the need for accurate and reliable precipitation time-series has increased. This has been caused, to a large degree, by the rising concern over global climate change and the need to detect changes in the spatial and temporal distribution of precipitation. An increased emphasis on hydrologic modelling, which requires accurate precipitation inputs, also contributes to a need for reliable precipitation time-series. Results from these models are useful in agricultural and economic applications, the detection of the impact of land-use changes, and the design of culverts and stream channels.

The accuracy of precipitation measurements, however, is undermined by two very important biases. First, inhomogeneities in the existing time-series may be caused by non-climatic factors such as changes in instrumentation and recording practices, temporal variations in siting characteristics, and station relocation. In addition, biases also may be induced by the gage measurement process through the deleterious effects of the wind, wetting on the interior walls of the gage, and evaporation from the gage. To evaluate climate change or the effect of changing land use over time, it is necessary to isolate the regional-scale precipitation fluctuations from the effects of local changes to the environment surrounding the gage. Thus, each of these concerns must be addressed if accurate and reliable precipitation time-series are to be obtained.

## Homogeneous Time-Series

Not all precipitation gages equally measure the amount of precipitation that falls. Gage catch can vary considerably with the use of different gage designs and different types of gage shields (cf., Legates 1987). For snowfall in particular, variations in gage catch can be as large as a factor of three for wind speeds of only five meters per second (Table 1). As gage designs have improved, many countries have adopted the new gages as a national standard which introduces a marked discontinuity into their precipitation time-series. For example, the Nipher-shielded gage which had long been used in the Soviet Union was replaced by the Tretyakov-shielded gage between 1948 and 1953. The United States adopted the use of Alter wind shields for some stations, particularly in the northwest, in the 1940s while Canada, Finland, Norway, Poland, Sweden, and Switzerland also have changed gage designs or shields in the past one hundred years (Groisman 1991, Groisman *et al.* 1991). These changes in gage design and shielding usually have resulted in an increased efficiency of the gage which, in turn, have caused in an increase in the gage catch. This translates to a sudden increase in the measured precipitation time-series which is not representative of the actual precipitation.

As gage catch usually decreases with increasing wind speed (Table 1) and since wind speed increases with height in the boundary layer, changes in the standard height of the gage orifice also adversely affect precipitation time-series (Neff 1977). Several countries have decreased the elevation of their gage orifice since the 1940s and in the United States, many gage installations have been moved from the roofs of downtown buildings to ground locations at suburban airports (Groisman *et al.* 1991). When these moves result in a change

in the mean wind speed over the gage orifice, a marked discontinuity is introduced into the precipitation time-series. Additionally, changes in the standard recording practice of snowfall also adversely affect the reliability of precipitation time-series (Goodison 1981).

**Table 1.** Mean catch (in percent) of several gages as a function of the mean wind speed at gage orifice height (from Goodison and Vet 1989 and interpolated from Koschmieder 1934).

Gage Type	Mean Wind Speed in Meters per second						
	1	2	3	4	5	6	7
<b>Rainfall</b>							
Alter-shielded Universal Gage	98	96	94	92	90	87	85
Unshielded Hellmann Gage	98	96	93	90	86	81	76
<b>Snowfall</b>							
Alter-shielded Universal Gage	88	77	67	59	51	45	40
Canadian Nipher- Shielded Snow Gage	104	106	106	102	96	87	77
Unshielded Universal Gage	63	53	45	38	32	27	23

Over time, it is unavoidable that some stations have been relocated for a variety of reasons. In particular, many of the United States National Weather Service Offices have been moved from urban locations to nearby airports situated in suburban or rural settings (Eischeid *et al.* 1991; Groisman 1991; Groisman *et al.* 1991). Owing to changes in the local environment, such relocations may introduce discontinuities into the station record. Separately or in combination, the effects of instrumental and observational changes and station relocation can yield a misleading picture of temporal climate changes in precipitation by introducing step discontinuities into the precipitation record. Consequently, they may be detected through a careful evaluation of the time-series (*e.g.*, Karl *et al.* 1990, Eischeid *et al.* 1991).

While searching for step discontinuities is useful, it does not necessarily ensure that a homogeneous precipitation time-series will result since the environment surrounding the gage may change considerably over time. For example, trees and bushes near the gage may grow or be cut down, buildings and fences near the gage may be constructed or demolished, or increased urbanization may occur. When the siting characteristics are altered, the wind flow around the gage will be affected which, in turn, may adversely influence the gage catch (*cf.* Brown and Peck 1962, Eischeid *et al.* 1991). Sudden changes caused by the construction of a nearby building, for example, may produce a detectable



step discontinuity into the time-series. Gradual changes over time such as tree growth may introduce a more gradual bias which can be difficult to detect by simply looking for discontinuities in the station record. Nevertheless, this bias may be difficult to distinguish from a true climate change signal and thus may be misleading.

Consequently, searching for and removing only noticeable discontinuities in the station record is insufficient to produce a truly homogeneous time-series. As a remedy, the use of station histories, or metadata, to provide documentation of and significant changes in gage exposure, gage type and shield (if used), orifice elevation, information from which surface roughness length can be estimated, obstacles to the wind flow (e.g., buildings and trees), and anemometer height for the gage site has been proposed (Groisman et al. 1991). While a potentially extensive documentation of siting characteristics is required, it is necessary to completely account for and remove homogeneity biases from the time-series. Since this procedure is more physically-based and less subjective, it is preferable to searching for jump discontinuities.

### **Gage Measurement Errors**

Systematic biases in gage measurement of precipitation may be introduced by the effect of the wind, wetting on the interior gage walls, evaporation from the gage, splashing into and out of the gage, blowing and drifting snow, and automatic recording techniques. The largest source of bias is the effect of the wind (Table 1) which induces a minor updraft, a small pressure difference between the air flowing over the gage and that in the orifice, and an slight increase in the wind speed over the gage orifice. These effects combine to decrease the gage catch and become more significant as the wind speed increases. Since many gages and shields vary considerably in design and may be used with different orifice elevations, the wind effect is gage specific.

During precipitation, moisture adheres to the funnel and is not measured in the collector. For gages which use a collector that must be emptied before a measurement can be made (e.g., the Tretyakov gage), wetting losses on the collector's interior surfaces additionally occur. Evaporation from non-recording gages occurring between the end of precipitation and its measurement also is a potential source of gage undercatch. While these losses can appear trivial, they represent a significant amount of moisture when integrated into monthly totals (cf., Legates 1987).

Splashing losses include moisture that is splashed from surrounding surfaces into the gage or splashed from the funnel out of the gage orifice. However, virtually all standard gage installations, with the exception of pit gages, site the orifice sufficiently above the ground so that splashing into the gage is avoided. As most gage designs employ a deep funnel, splashing out of the gage is insignificant. Compared to the effects of the wind, wetting, and evaporation, splashing errors are negligible except when pit gages are employed.

Fallen snow may be lifted into the air during periods of high winds when a significant amount of dry snow exists on the ground. This blowing snow can be trapped by some gage shield designs which results in an erroneous increase in measured precipitation. This is a particular problem for the Tretyakov-shielded gage (Struzer and Bryazgin 1971) although it occurs mainly in higher latitudes and high elevations.

Automatic recording techniques also are responsible for systematic underestimates of precipitation. For example, the weighing mechanism of weighing gages, floats in float gages, and the recording pen unit of many recording gages are affected to some degree by a

slight frictional drag which can result in a small decrease in the measured precipitation. In addition, precipitation may also be underestimated by some self-emptying gages during the time it takes the gage to empty (Linsley *et al.* 1982).

Due to their widespread use, tipping bucket raingages are probably the biggest source of mechanically-induced biases. Small quantities of moisture can be lost during tips of the bucket and this problem is particularly accentuated during periods of intense rainfall. In many instances, the precipitation recorded by the number of bucket tips can be compared with the precipitation measured in the bottom of the collector and the discrepancy assigned to the high intensity periods (Parsons 1941). For measurement of snowfall, electrically-heated tipping bucket gages may be used although this enhances the sublimation of newly-fallen snow and evaporation of melted snow. Thus, snowfall measurements using heated tipping bucket gages may be unreliable. Although small, mechanical errors nevertheless result in a systematic decrease in the measured precipitation.

Beginning with Sevruk's (1979) general model for precipitation correction and following Legates (1987), an estimate of the "corrected" (unbiased) precipitation estimate,  $P_c$ , can be obtained from

$$P_c = k_r ( P_{gr} + sP_{wr} + sP_{er} + sP_{mr} ) \\ + k_s ( P_{gs} + sP_{ws} + sP_{es} + sP_{ms} ) \pm sP_b$$

where  $k$  is the wind deformation coefficient ( $k \geq 1$ ) which increases as the wind speed increases,  $P_g$  is the gage measured precipitation,  $sP_w$  is the wetting loss,  $sP_e$  is the evaporative loss,  $sP_m$  is the correction for mechanical effects,  $sP_b$  is the correction for blowing snow, and the additional subscripts,  $r$  and  $s$ , denote liquid (rain or drizzle) and solid (snow) components, respectively. Unbiased estimates of precipitation can therefore be obtained by evaluating each of the terms in this equation which are both gage and site specific (Legates 1987). Since the correction cannot be perfect and will introduce some degree of statistical error, it is imperative that with the unbiased estimate, a confidence interval or a measure of the standard error of the correction be provided.

### **The Influence of Variations in Other Meteorological Variables**

Over the last couple of decades, a number of large-scale terrestrial precipitation climatologies have been compiled. Some have provided time-series that have been scanned for significant jump discontinuities (*e.g.*, Bradley *et al.* 1987, Diaz *et al.* 1989, Karl *et al.* 1990, Vinnikov *et al.* 1990, Eischeid *et al.* 1991) while others have focussed on mean monthly estimates from which gage measurement biases have been removed (*e.g.*, UNESCO 1978, Legates 1987, Legates and Willmott 1990). However, it is apparent that both issues -- station heterogeneity and biased estimates -- must be considered to obtain an accurate and reliable precipitation climatology. This is particularly true since changes in other meteorological variables can affect the precipitation time-series.

Consider, for example, an Alter-shielded Universal gage sited in an area where significant snowfall occurs during the winter and air temperatures rise well above freezing in the summer. Assume further that a completely homogeneous time-series exists for a number of years; that is, no changes in instrumentation, siting characteristics, *etc.* occurred. Owing to natural intra-annual variations, it is expected that mean monthly air temperatures and wind speeds would fluctuate from year to year. Thus, some years will be

warmer while others will be colder which, during spring and autumn, should vary the frequency of snowfall. However, even if actual precipitation remained the same from year to year and did not fluctuate, the measured precipitation record would still exhibit yearly fluctuations since variations in air temperature and wind speed would differentially affect the catch for each year.

This is illustrated in Figure 1. Goodison and Vet (1989) provide the mean gage catch as a function of mean wind speed for the Alter-shielded Universal gage while the logistic regression of Legates (1987) estimates the proportion of snowfall to total precipitation as a function of the mean monthly air temperature. Variations in mean wind speed influence the gage catch to a much greater degree when the precipitation falls primarily as snow (colder temperatures). Note that the gage is shielded -- the effect would be even more pronounced with an unshielded gage. In addition, variations in air temperature exert a greater influence on the gage catch at air temperatures just a couple of degrees below freezing. Here, the logistic curve has its steepest slope and indicates that a small change in air temperature can greatly affect the ratio of monthly snowfall to total precipitation.

Consider again this hypothetical station. If urbanization occurs during the period of record, it would not be unreasonable to assume that the mean wind speed might decrease slightly while the mean air temperature increases. A decrease in wind speed would result in a more efficient measurement of precipitation. During spring and autumn, the increase in air temperature would decrease the snowfall and increase rainfall frequencies. Since the gage is more efficient in measuring rain than snow, the observed precipitation record would exhibit a slight increase over time as a result of these two effects. Urbanization, however, need not be the cause for the changes in air temperature and wind speed as it could occur from natural variability or a variety of other climate change mechanisms.

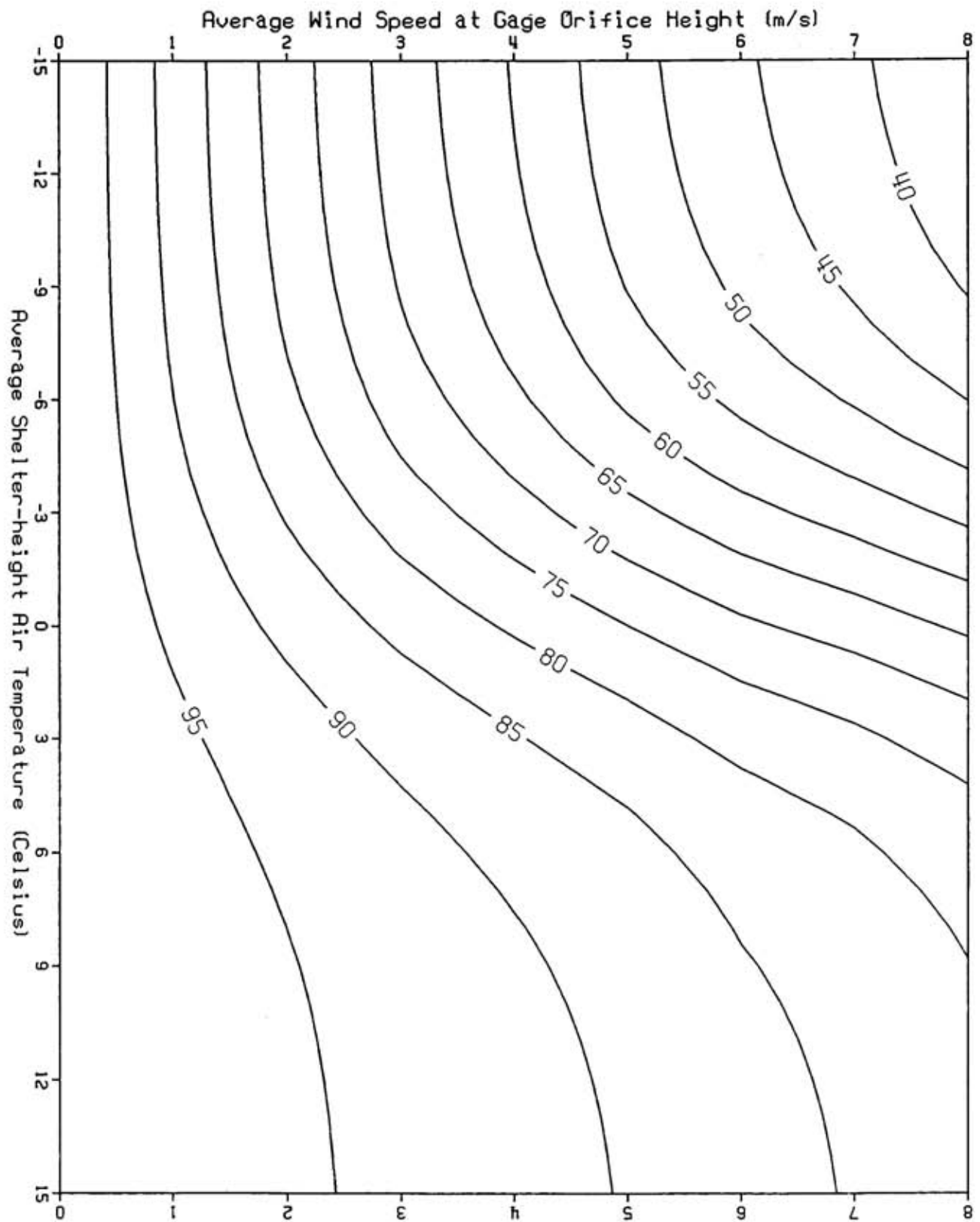
## **Recommendations**

A holistic approach to obtaining more accurate and reliable precipitation time-series is accomplished by coupling corrections for gage measurement biases with the removal of non-climatically induced station inhomogeneities. Many of the procedures and equations discussed by Sevruk (1982), Legates (1987), and in an edited volume (Sevruk 1986) explicitly account for gage type and siting characteristics. Through these equations, most of the inhomogeneities caused by changes in instrumentation, recording practices, station relocation, and siting changes can be removed while simultaneously providing more accurate estimates. Only by explicitly accounting for both of these potential sources of error can a truly unbiased and homogeneous precipitation time-series be realized.

These procedures are further being refined and improved through the WMO Solid Precipitation Measurement Intercomparison (Goodison *et al.* 1988, Goodison *et al.* 1989). In this project, snowfall measurements from a number of national standard gages are being compared to the double-fence instrument reference (DFIR) at approximately twenty-four sites around the world. When complete, this project will provide a more complete assessment of the influence of wind speed on a variety of gages. It is hoped that through this and similar efforts, the accuracy, reliability, and compatibility of precipitation time-series will be increased.

## **Acknowledgements**

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**Figure 1.** Ratio of the mean gage catch to the actual precipitation (in percent) as a function of average wind speed and air temperature for an Alter-shielded Universal gage. Mean gage catch for rain and snow events is given by Goodison and Vet (1989) and the proportion of solid precipitation as a function of air temperature is taken from a logistic regression equation developed by Legates (1987). Wetting and evaporative losses were not considered.



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# Comparison of Two Satellite-Derived Albedo Data Sets for the Arctic Ocean

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## Abstract

Spatial patterns of parameterized mean monthly surface albedo derived from DMSP satellite imagery are compared with surface albedos derived from the International Satellite Cloud Climatology Program (ISCCP) monthly data set. Spatial patterns of albedo obtained by the two techniques are generally in good agreement with each other, especially for June, July and August. Nevertheless, systematic differences in albedo of from 0.05 to 0.10 are noted. These differences are most likely related to problems in the ISCCP cloud-clearing algorithm and other modelling simplifications, as well as uncertainties in the simple parameterizations used in the DMSP analyses. However, with respect to the eventual goal of developing a reliable automated retrieval algorithm for developing a long-term albedo data base, these initial comparisons are very encouraging.

## Introduction

Variations in the surface albedo of the Arctic pack ice cover may be an important forcing factor of northern high-latitude climates, with implications for the long-term mass balance and stability of the pack ice and potential impacts on other parts of the Northern Hemisphere [Fletcher, 1966; Barry, 1983]. Continued monitoring of albedo is consequently an important component of the "early detection" strategy for identifying global and regional climate changes [Barry, 1985]. Nevertheless, until recently, studies of the Arctic Ocean albedo have been of limited spatial and temporal scope, based largely on observations at drifting stations, on fast ice, and during aircraft missions. Other investigators have used these measurements and satellite passive microwave data to estimate regional summer albedos [see Robinson et al., 1992 and references therein].

With the aim of developing a long-term record, Robinson et al. [1992] recently developed a ten year data set of parameterized albedo for the entire Arctic Ocean for May through mid-August. Maps of surface brightness, charted through manual analysis of visible-band satellite imagery, were converted to parameterized surface albedo using published data and image processor techniques. Although lacking radiometric control, this data base (hereafter referred to as the RO92 set) has allowed for definition of the climatological characteristics of albedo over the entire Arctic Basin. More recently, we have used data from the International Satellite Cloud Climatology Program (ISCCP) monthly (C2) data set in conjunction with modeled snow albedos to automate basin-wide retrieval of surface albedo (hereafter referred to as the ISCCP data set). Here, we make some initial comparisons between these two data sets. While providing a useful evaluation of uncertainties in our current knowledge of Arctic Ocean surface albedo, this paper also addresses problems in the automated extraction of these data, and clarifies aspects of the existing retrieval algorithm in need of further attention.

## Data and Methods

### *DMSP Data Set*

The RO92 data set has its basis in a manual charting of surface brightness changes, in three day increments over the entire Arctic Ocean for May-mid June of 1975, 1977-1980 and 1984-1988. Analyses are based primarily on 2.7 km resolution Defense Meteorological Satellite Program (DMSP) orbital swath images. These data, which provide daily Arctic-wide coverage, are archived as transparencies only, produced operationally from digital data in a broad-band channel spanning the visible and near-infrared (0.4-1.1  $\mu\text{m}$ ) wavelengths. DMSP images with local coverage at a 0.6 km resolution, available for the Beaufort and Chukchi seas only, and NOAA Advanced Very High Resolution Radiometer (AVHRR) 1.1 km resolution visible-band images, were used when the 2.7 km DMSP products were missing from the archive or of poor quality [see Scharfen et al., 1987 and Robinson et al. 1992 for details].

The brightness changes observed in the images are primarily due to snow melt, exposure of bare ice, the formation and drainage of melt ponds [Barry, 1983] and, to a lesser extent, changes in ice concentration [Robinson et al., 1992]. Efforts were made to chart the brightness for what appears on the images as snow/ice, ignoring the effects of large leads and polynyas. Four brightness classes were defined [Scharfen et al., 1987; Robinson et al., 1992]. Available surface and aircraft observations indicate that brightness class 1 corresponds to fresh snow cover over 95% of the ice; class 2 is found when snow covers between 50-95% of the surface, with the remainder being bare or ponded ice; class 3 represents the advanced to final stage of snow melt, with numerous melt ponds and between 10-50% of the ice surface snow covered, or, following pond drainage, predominantly bare ice; class 4 is heavily-ponded or flooded ice, generally limited to regions of fast ice near outlets of major rivers along the Siberian Arctic coast.

Typically 15-25 images were used during each charting interval, which was usually found to be long enough to obtain at least one clear-sky image over a given area yet short enough to permit evaluation of temporal and regional variations in surface brightness. Surface/cloud discrimination was based on identification of cloud shadows, cloud motion and characteristic features of the surface including ice floes, leads and melt ponds [Robinson et al., 1985; Serreze and Rehder, 1990]. Extensive cloud cover and poor lighting precluded charting after mid-August. The surface brightness maps were digitized to the Limited-Area Fine Mesh version of the U.S. National Meteorological Center (NMC) grid, dividing the Arctic Ocean into 223 cells. Grid cells were simply assigned the value of the predominant brightness class in that cell, with open-water cells defined using Navy/NOAA Joint Ice Center (JIC) ice concentration charts (based primarily on satellite data; see Godin, 1981) closest in time to the analyzed three-day interval, digitized to the same NMC grid. Generally, 20% of all ice-covered cells were initially missing from each chart, primarily due to cloud cover. Cells with missing data were assigned the values of corresponding cells from the immediately preceding chart or subsequent chart if available. Data were usually available over more than 90% of all cells for each chart.

Parameterized albedos were next assigned to each grid cell. Digital numbers (DNs) were measured for targets of sea ice and open water within rectangular regions of approximately  $9000 \text{ km}^2$  from clear-sky portions of 20 digitized DMSP transparencies for May through July of 1977 and 1979. Both the 0.6 km and 2.7 km DMSP products were used. A total of 158 rectangular regions were examined, with approximately 1400 DN values obtained for each region. The highest mean DN values of snow-covered sea ice and the lowest mean DN of open water were then used as tie points and assigned clear-sky

albedos based on measured ground and aerial data. The upper tie-point brightness was assigned an albedo of 0.79 until late June, after which a value of 0.69 was used, due to the decrease in maximum image brightness associated with snow melt [Robinson et al., 1992]. The albedo of open water cells was taken as 0.12 [Cogley, 1979]. The literature values of albedo, and hence our resultant values, refer to the integrated solar spectral range of approximately 0.3-2.7  $\mu\text{m}$ .

The average DNs from regions corresponding to the charted classes were then converted into clear-sky albedos by linear interpolation between the tie points. The resulting mean clear-sky values were 0.75 (class 1), 0.59 (class 2), 0.44 (class 3) and 0.27 (class 4), with standard deviations of 0.04 (class 1), 0.07 (class 2), 0.08 (class 3) and 0.05 (class 4). Cloud cover tends to increase albedo due to preferential absorption of near-infrared radiation and multiple reflections between the surface and cloud base, the latter effect being minor when surface albedos are low. To obtain albedo values representative of typical Arctic cloud cover conditions, the clear-sky albedos were increased by 0.05 for brightness classes 1-3 and 0.02 for class 4, based on published data [see Robinson et al., 1992] with a constant 75% cloud cover assumed for each grid cell [Kukla and Robinson, 1988].

On the basis of width distributions of sea ice leads (cracks) examined from submarine sonar data [e.g., Key and Peckham, 1991] with respect to those detectable in the DMSP imagery, these class albedos are considered to be weighted by open water in the pack ice. The effects of undetected areas of dark, new thin ice will also be included in these estimates. Again, the reader is referred to Robinson et al. [1992] for further details.

Clearly, these albedo data are only appropriate for broad climatological analysis. First, each melt class corresponds to a broad range in albedo. Second, while in support of the simple treatment of cloud cover, monthly maps for summer by Gorshkov [1983] and Serreze and Rehder [1990] show little spatial variation in cloud cover over most of the pack ice (primarily low-level Arctic stratus, see Herman and Goody, 1976), there can be large interannual and short-term variability [Barry et al., 1987; Serreze and Rehder, 1990; Robinson et al., 1985]. Third, since the effects of large, visible leads and polynyas were excluded during charting, the derived albedos will tend to be slightly high. Furthermore, no account is taken of spatial or temporal variations in these features.

### *ISCCP Albedos*

The International Satellite Cloud Climatology Project (ISCCP) monthly (C2) data set [Rossow and Schiffer, 1991] contains cloud amount, cloud optical depth, 0.6  $\mu\text{m}$  surface reflectivity, surface temperature, as well as atmospheric profiles of temperature, water vapor and ozone. These parameters are almost exclusively derived from satellite. By design, the ISCCP data set is suitable for the calculation of radiative fluxes at the surface and the top of the atmosphere. The ISCCP C2 data set is a compilation of monthly statistics from the 3 hourly C1 data set. While the ISCCP data have recently been expanded to cover the period July 1983 through 1989, only data through 1986 are examined here. Additionally, we restrict our analyses to ocean areas north of 62.5° latitude.

The spatially integrated albedo of snow or ice covered surfaces depends not only on the spectral albedo of the surface but also on the spectral distribution of the downwelling shortwave radiation. Since the ISCCP data set only reports reflectivities at 0.6  $\mu\text{m}$  (AVHRR channel 1), radiative transfer calculations were conducted in order to obtain



spectrally integrated albedos in a waveband comparable to those reported by Robinson et al. [1992].

The radiative transfer model used to calculate radiative fluxes in this study is that of Tsay et al. [1988]. Gas absorption for water vapor, ozone, carbon dioxide and oxygen is parameterized using an exponential sum fitting technique [Wiscombe and Evans, 1977] with 24 bands at varying intervals without overlap. Atmospheric temperatures reported in the ISCCP data set were supplemented with inversion statistics from ice islands [Serreze et al., 1992] for more representative temperature profiles. Approximately 10 vertical layers are used, with the temperature gradient across each being less than 10 K in order to satisfy the limitations of the linear approximation of the Planck function in the presence of clouds. Clouds are modeled as Mie scattering layers of variable thickness, with physical height and optical thickness given in the ISCCP data set. Cloud physical thickness is calculated assuming an effective radius and liquid water concentration consistent with the ISCCP retrieval algorithm.

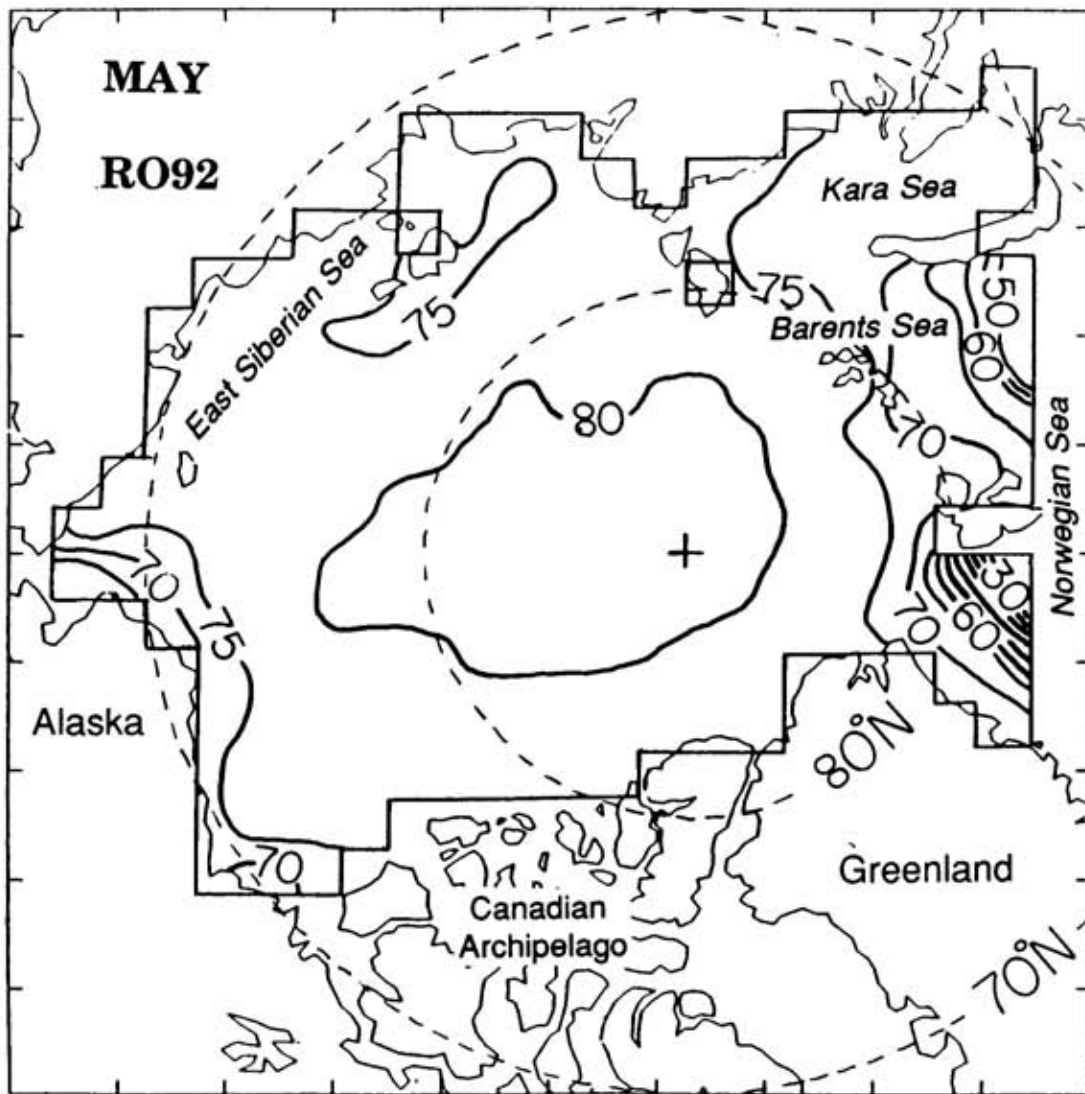
Since the ISCCP data set only provides a single channel (0.6  $\mu\text{m}$ ) measurement of surface reflectivity, reflectivities for the remaining spectral bands in the radiative transfer scheme have to be inferred from this measurement based on surface type(s). The surface is assumed to be a mixture of no more than three types, selected depending on season from a catalog consisting of fresh snow, old melting snow, melt ponds, bare ice, and open water. For example, in June, when the snow cover is melting, the prescribed surface types are ice, melt ponds and open water. Given the channel 1 reflectivity for the entire cell, the ice concentration contained in the ISCCP data set (from the JIC charts) and expected 0.6  $\mu\text{m}$  reflectivities for each of the assumed surface types, the fractional area of melt ponds can be calculated. If melt pond coverage is calculated to be an unrealistic value (e.g., if it is present when only snow and open water are assumed to exist or when the coverage is greater than 50% otherwise), channel 1 snow reflectivities are adjusted to match observed values.

Reflectivities for 24 shortwave bands (total range of 0.28-4.0  $\mu\text{m}$ ) were computed for the two snow types assuming Mie scattering of ice grains and soot particles [Warren and Wiscombe, 1980, Wiscombe and Warren, 1980]. Spectral reflectivities of melt ponds and bare ice were taken from Grenfell and Maykut [1977], with open water reflectivities taken from Briegleb et al. [1986]. Snow reflectivities in the remaining 23 bands are determined under the assumption of constant ratios to channel 1 reflectivities. The fractional coverage of each surface type is then used to calculate the total surface reflectivity for the entire range of the 24 bands.

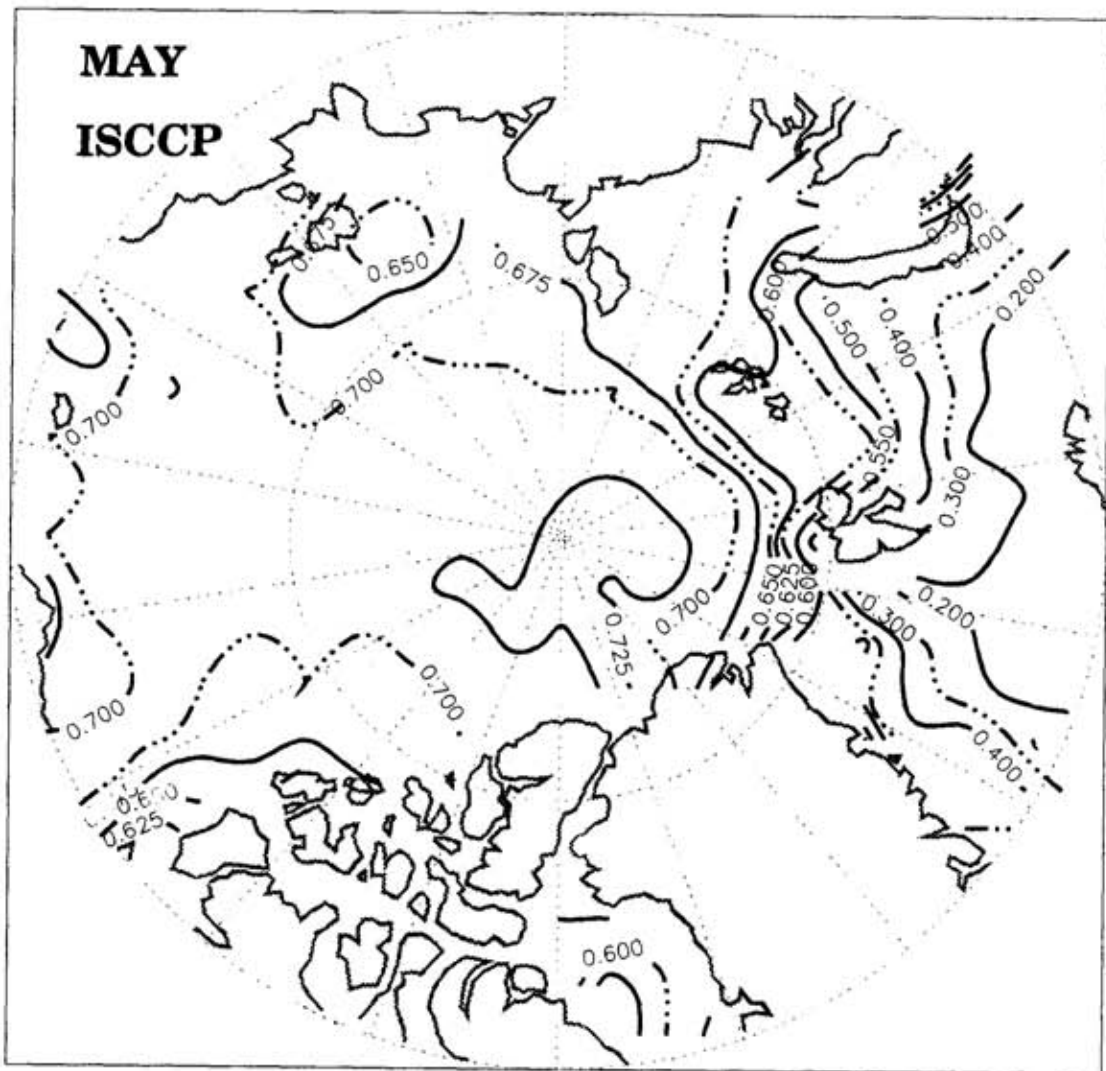
## Results

Figures 1 through 4 show the spatial patterns of mean monthly surface albedo from the RO92 and ISCCP data sets for May through August. During May, when snow melt is largely confined to coastal regions [Robinson et al., 1992] both data sets show little spatial variability over the central Arctic. The RO92 albedos over the central Arctic tend to be between 0.75 and 0.80 (the latter figure representing the maximum possible parameterized value). By contrast, the ISCCP values over the same regions are lower by typically 0.050 to 0.075. Both data sets are in agreement in showing a strong gradient of decreasing albedo near coastal regions, especially toward the Norwegian and East Siberian seas, consistent with the effects of decreasing ice concentrations [Godin, 1981] as well as coastal snow melt [Robinson et al., 1992]. Both data sets also show surface albedos to be lower near the New Siberian Islands. The reduction in albedo near the Canadian Arctic Archipelago shown by ISCCP, however, is not depicted in the RO92 analysis.





**Figure 1a.** Mean monthly albedo for May from the RO92 data sets.



**Figure 1b.** Mean monthly albedo for May from the ISCCP C2 data sets.

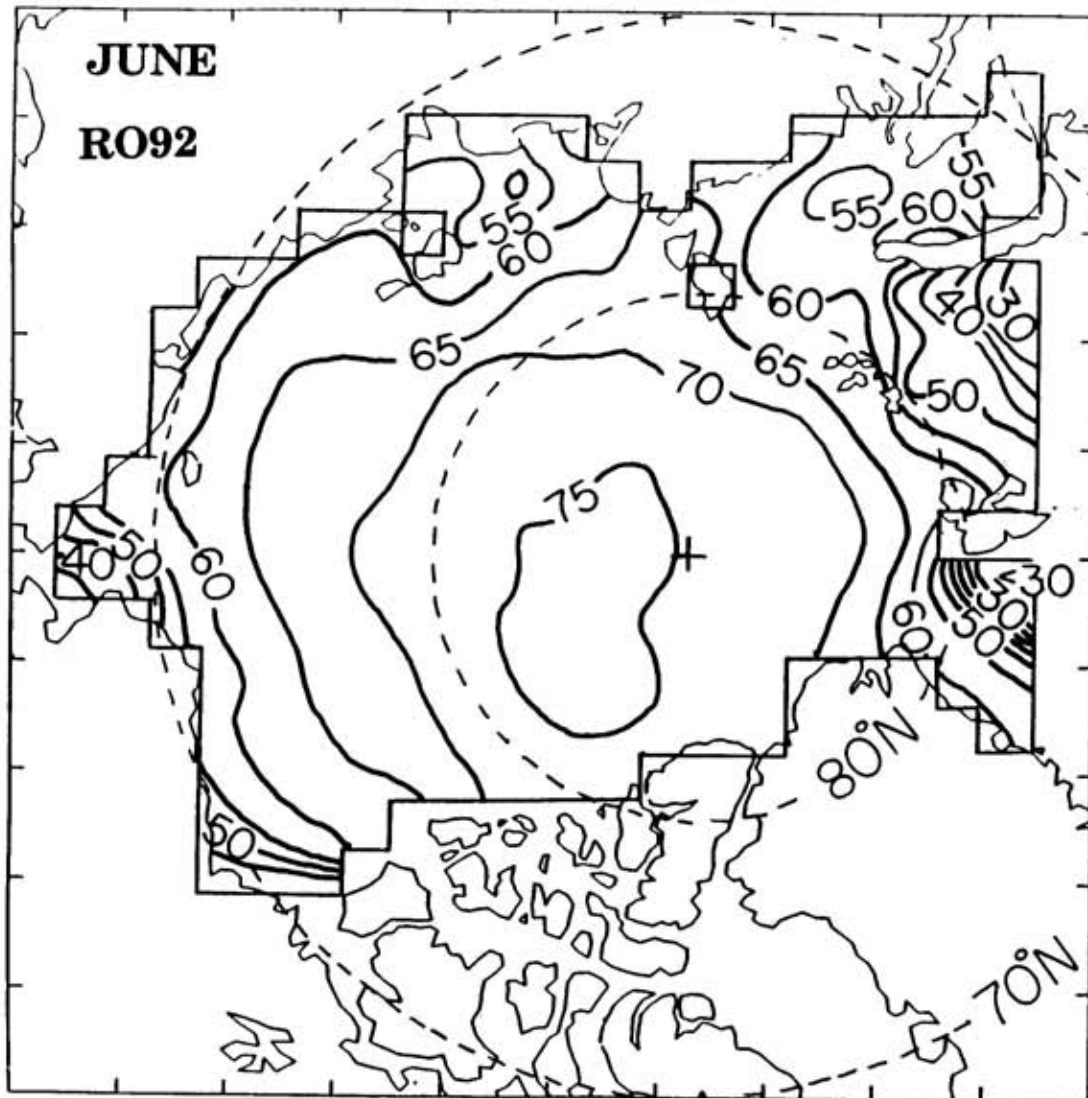
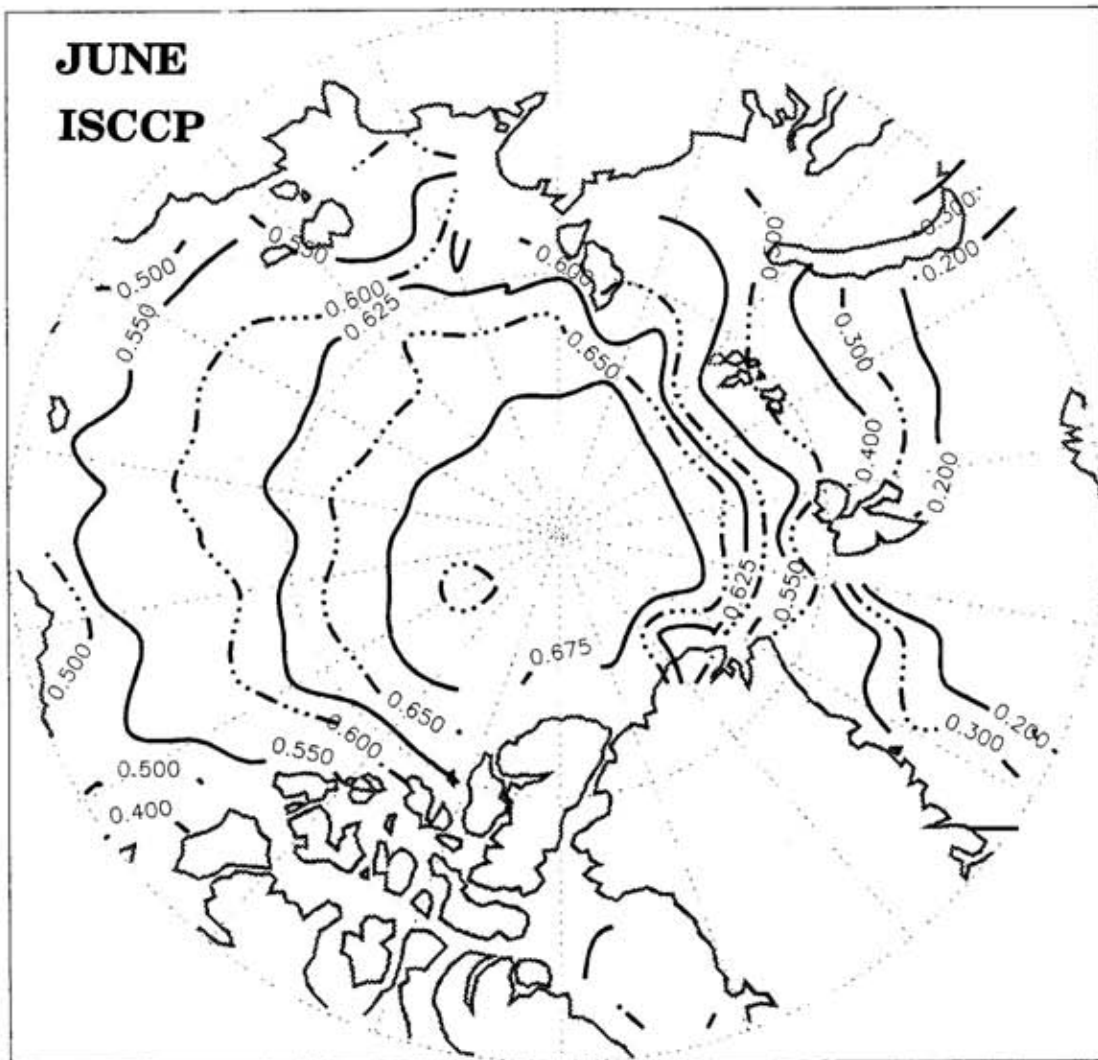
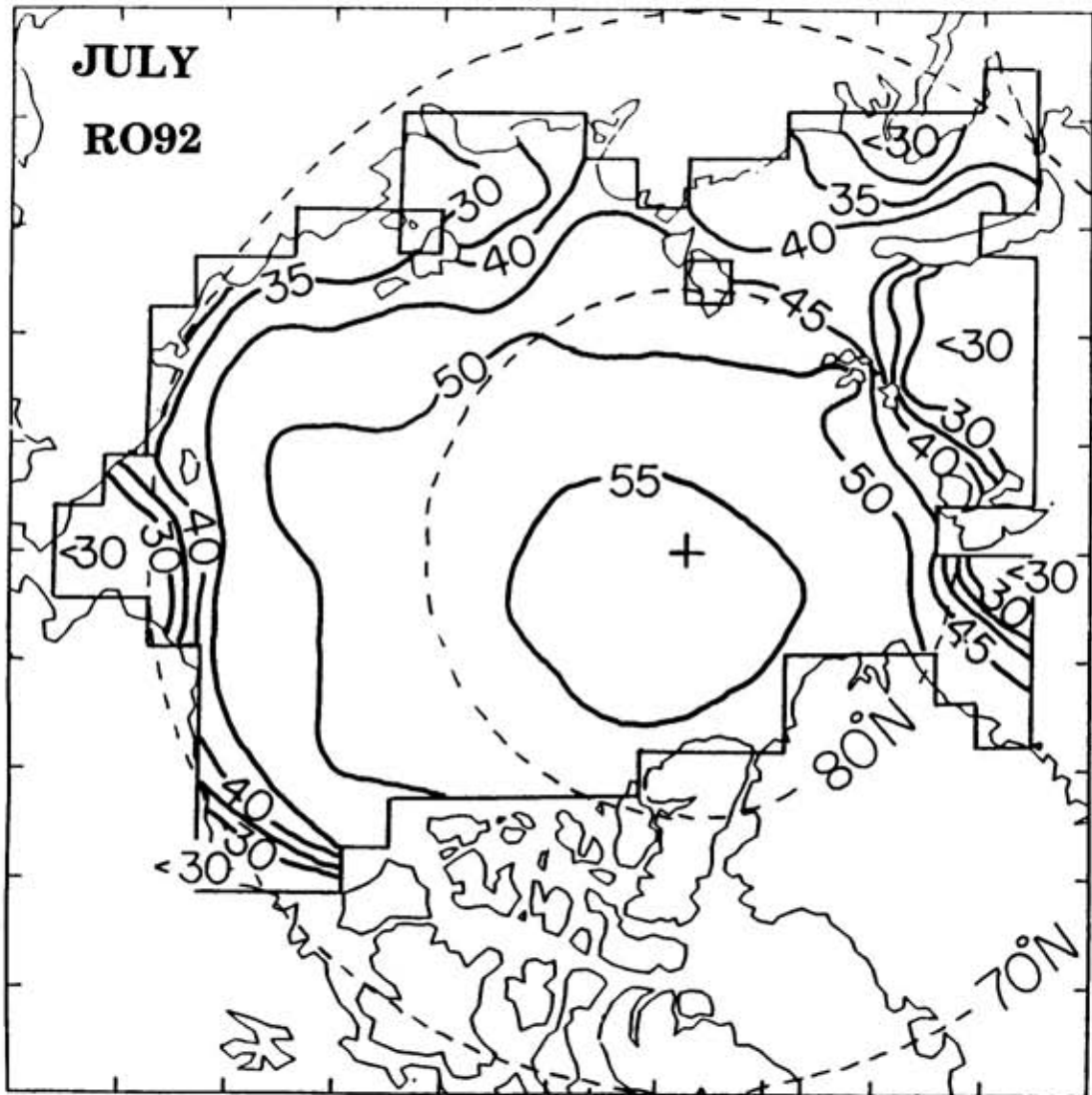


Figure 2a. Mean monthly albedo for June from the RO92 data sets.

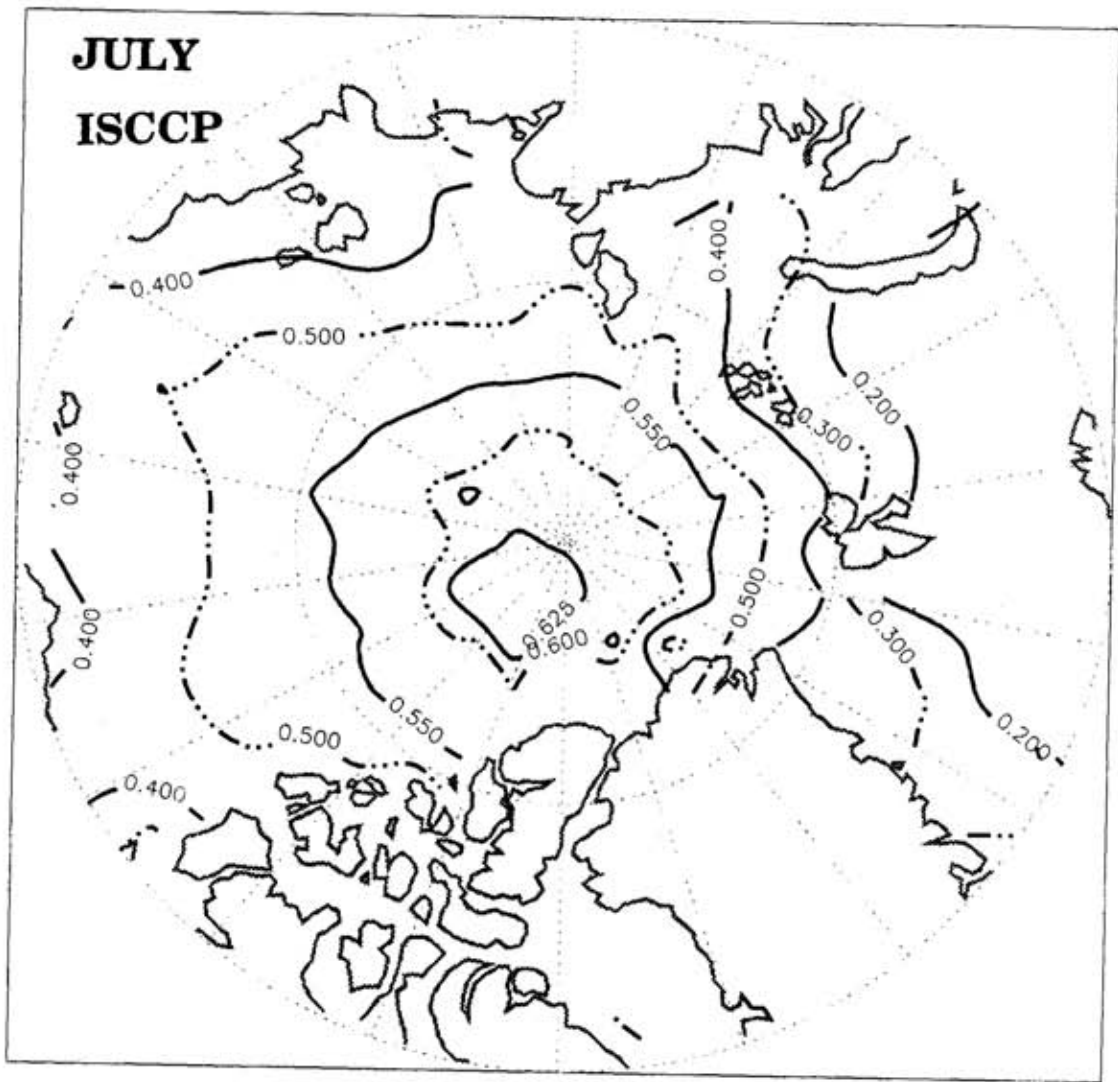


**Figure 2b.** Mean monthly albedo for June from the ISCCP C2 data sets.

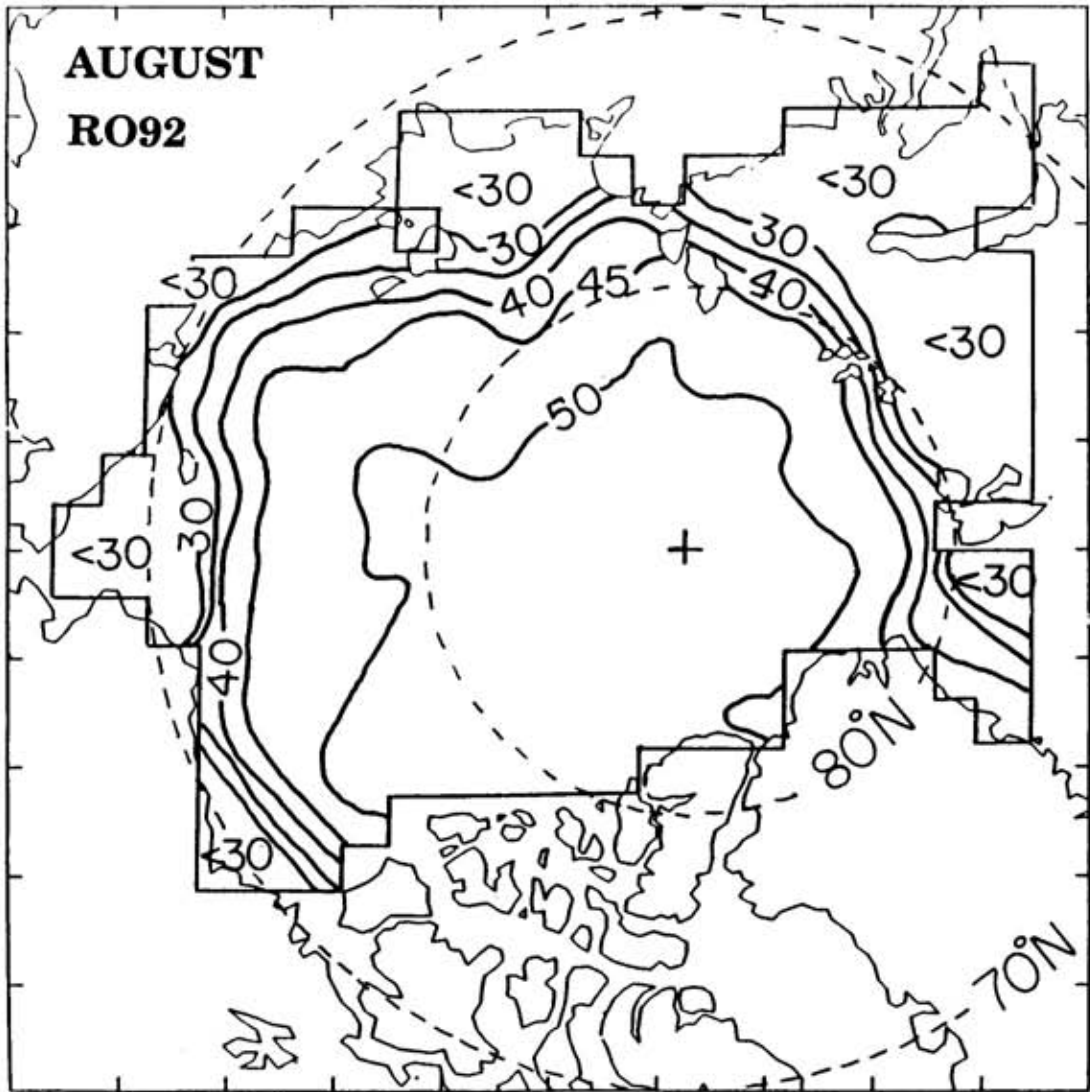


**Figure 3a.** Mean monthly albedo for July from the RO92 data sets.

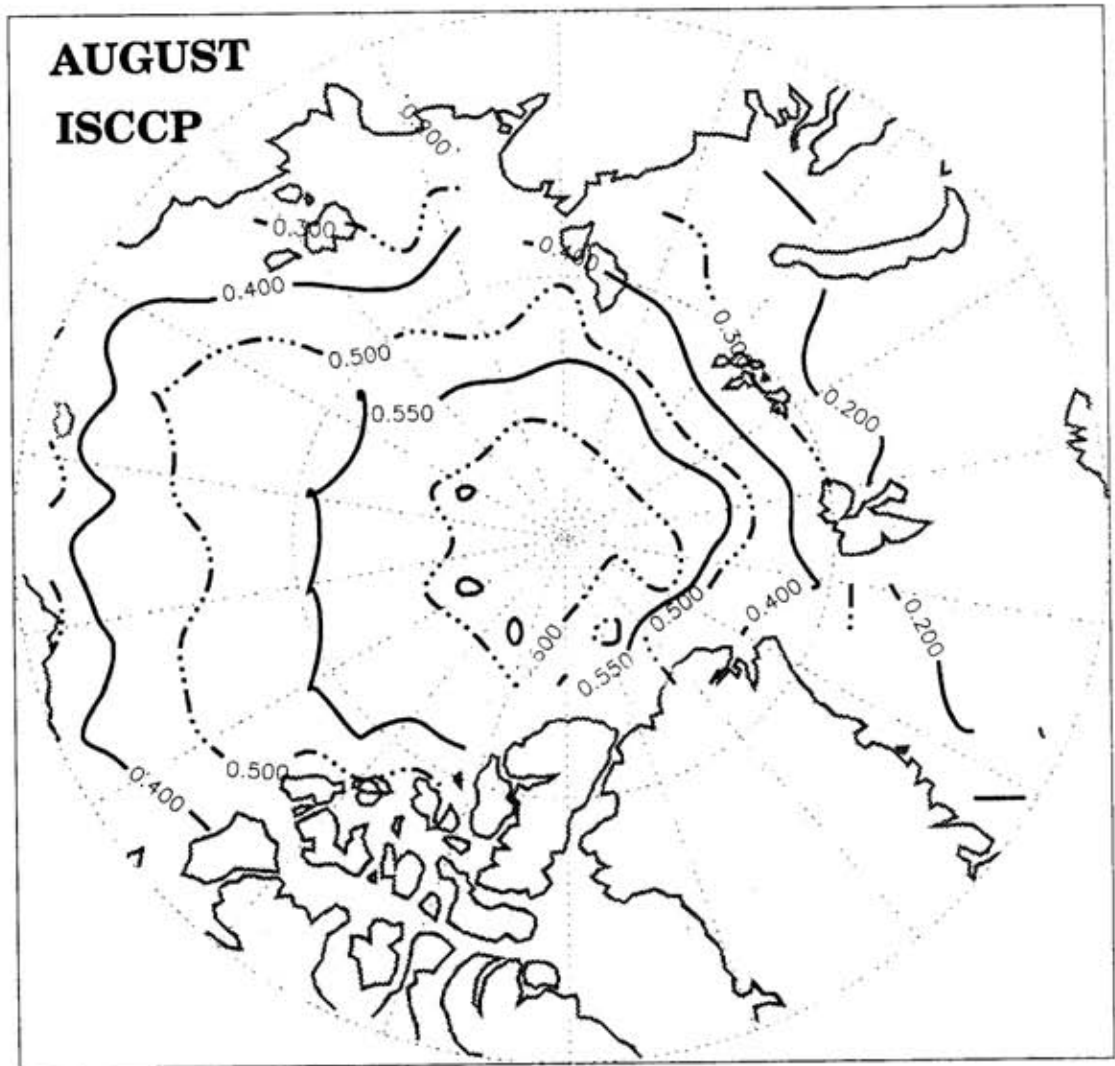




**Figure 3b.** Mean monthly albedo for July from the ISCCP C2 data sets.



**Figure 4a.** Mean monthly albedo for August from the RO92 data sets.



**Figure 4b.** Mean monthly albedo for August from the ISCCP C2 data sets.

By June, melt is underway over large areas of the sea ice cover [Robinson et al., 1992], which results in a decrease in surface albedo (Figure 2). Both data sets show a concentric pattern of surface albedos, with values decreasing from the pole southwards. Serreze et al. [1991] show that this concentric pattern can be related to the climatological distribution of melting degree days. ISCCP albedos, however, again tend to be lower than those reported by RO92. For example, in the central Arctic regions ISCCP tends to be lower by about 0.075.

In July, albedos continue to drop from their June values (Figure 3). The concentric pattern observed in both data sets for June is still present. During this month, the ISCCP and RO92 albedos match very well, except near the pole where ISCCP albedos are now higher than the RO92 values by up to 0.075. In August albedos have dropped further (Figure 4) but, as with July, the ISCCP albedos in the central Arctic are higher than those shown by RO92.

## Discussion

The RO92 ISCCP data sets are in good agreement with respect to spatial patterns, especially for June through August. These spatial patterns also compare favorably with the summary of surface melt prepared by Marshunova and Chernigovskiy [1978], in which a concentric pattern of melt over the Arctic pack ice progressing towards the pole by early July is evident (not shown).

One of our goals is to establish a long term Arctic surface albedo data set, using an automated retrieval scheme such as the ISCCP analysis procedure. In this regard, the generally good agreement between the ISCCP and RO92 data sets, which are calculated entirely independently, is very encouraging. However there are systematic differences in the absolute values reported in the two data sets on the order of 0.10. Specifically, the ISCCP albedos during May and June are less than those reported by RO92, but are larger during July and August near the Pole. The August difference can be attributed, at least in part, to the fact that RO92 values are based on the first half of the month (August 1-15), whereas the ISCCP values represent the entire month. During the second half of August new snow is often present so that the monthly mean albedo is expected to be higher. There is probably no single source for the other discrepancies.

As for likely candidates, it may be significant that the ISCCP algorithm in its current version assumes that clouds have a higher channel 1 reflectivity than the surface. In the polar regions this may not be the case. Because of the conservative nature of the ISCCP algorithm such ambiguous cases would be discarded, thereby acting to bias the ISCCP results towards lower surface albedos. On the same basis, the ISCCP algorithm appears to underestimate cloud amounts significantly [Schweiger and Key, 1992] and since the spectrally integrated albedo of snow covered surfaces under cloudy conditions can exceed those of clear sky conditions by more than 0.1, an underestimate of total cloud amount may also explain why ISCCP albedos are too low. Furthermore, in the calculation of spectrally integrated albedos from the ISCCP data set, a number of assumptions are made regarding the spectral albedos and distribution of surface types; e.g., grain size and soot content of the snow. However, since the ratio of visible band to near infrared reflectivity of snow is chiefly determined by those factors, spectrally integrated albedos for clear sky conditions are rather sensitive to their specification.

With regard to the RO92 analyses, one may argue that a human interpreter should provide a more accurate cloud/snow discrimination. This data does contain an adjustment of albedo by climatological cloud amounts but, given its crudity, this may further contribute

to the differences between the two data sets. More generally, the fundamental limitation of the RO92 set is that it lacks radiometric control. Only four albedo classes are provided, each with a large standard deviation. Angular and atmospheric effects are essentially ignored, as are the effects of nonlinearities in the original photographic recording process. It is interesting to note that if one were to remove the cloud cover adjustment from the RO92 data, the results would tend to become more in line with the ISCCP clear sky estimates, at least for May and June. This may be simply fortuitous, however, since it does not explain the larger ISCCP values during July.

Finally, one must also consider that since the ISCCP data are for four years only (1983-1986), compared to the 10-year RO92 data set, part of the problem may also lie in the natural interannual variability of albedo. The RO92 data set is not considered appropriate for addressing this interannual variability [Robinson et al., 1992], especially with respect to the use of only four class albedos. As such, we leave this point as an open issue.

## Conclusions

Good agreement in the spatial patterns of surface albedo is found between RO92 and ISCCP data sets. While the RO92 data set cannot be considered "ground truth", the similarity between albedo patterns in these two independently-derived products gives a positive outlook for operational retrieval of albedo from satellite data. However, there are systematic differences in the absolute albedo values, with ISCCP albedos being lower during May-June and higher in July-August. Reasons for these differences can be found in the methodology used to produce both data sets: the present inadequacies in the ISCCP cloud-clearing algorithm and the simplifications and uncertainties in the RO92 procedures. More work is needed to resolve these uncertainties before an accurate assessment of the surface energy balance can be made.

## Acknowledgements

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# Modelling The Interactions Between Microwaves And A Snow/Soil System

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## Abstract

A snow/soil thermal/backscattering model has been developed to study the various mechanisms driving the radar backscattering of a snow covered terrain. More specifically, the influence of the unfrozen water content in a frozen soil was taken into account in the calculation of the surface backscattering at the soil/snow interface. Backscatter simulations of snow covered loam, till, and clay soils have revealed significant positive relationships between  $T_0$  and snow depth, suggesting that, under the assumptions used to develop the model and in the simulation runs, snow depth and snow water equivalent can be extracted from radar data. It was shown that the observed relationships were best explained by the large thermal resistivity of snow, which affects the amount of unfrozen water in a frozen soil, thereby altering the soil dielectric constant and the radar backscattering coefficient at the snow/soil interface.

## Introduction

Seasonal snow is a major component of the hydrological cycle in temperate, alpine, and boreal environments. Snow cover properties, such as the snow water equivalent (SWE) and the areal extent of the snow pack, have been identified as among the most important parameters for flood forecasting, hydroelectric reservoir operation, and in water supply management. Moreover, snow cover plays a significant interactive role in the regional and global climate system, as it alters the heat and moisture exchange at the earth surface. Quantification of SWE and monitoring seasonal fluctuations in the snow cover extent are therefore required to monitor short changes as they impact the hydrological and energy cycles; and to improve our understanding of the snow-climate interactions and feedback mechanisms and evaluating the potential impacts of climate change on snow cover.

Previous research using passive microwave data from NIMBUS-7 SMMR (Kunzi, et al., 1982; Chang, et al., 1987) and DMSP SSM/I (Goodison, et al., 1989) sensors has shown the feasibility of deriving SWE and snow cover extent. However, the rather poor resolution of these sensors (typically 10's of kms) have limited their use to large basins, and the sub-pixel distribution of the snow cover would be difficult to retrieve from passive microwave brightness temperatures. Synthetic aperture radar (SAR) sensors, on the other hand, offer the advantage of very fine resolution (10's of meters), which is particularly attractive for monitoring snow cover properties in small watersheds or in small open areas. SAR signatures however, are influenced by numerous factors, for example roughness, incidence angle, moisture content, and topography, which makes the retrieving of snow parameters a non trivial task. In addition, dry snow is difficult to detect at the typical SAR microwave frequencies, which further complicates the task of extracting SWE.

In this paper, the potential of the radar for monitoring the snow depth/SWE has been investigated using modelling techniques. To that end, a coupled snow/soil backscatter/thermal model has been developed and used in sensitivity studies to assess the relative importance of snow and soil physical properties and condition on the radar backscatter of a snow covered surface. Properties include: snow depth, density, and soil surface roughness. Condition includes soil moisture, and whether the soil is frozen or not.

## Background

Research in the use of active microwave sensors for SWE extraction has been limited to a few ground-based scatterometer experiments (Ulaby and Stiles, 1980; Matzler and Schanda, 1984; Lowings et al., 1991) and airborne missions (Goodison et al., 1980; Leconte and Pultz, 1990; Leconte et al., 1990; Bernier, 1991). With the launch of ERS-1, national "Announcement of Opportunity" initiatives will allow researchers to evaluate the possibilities and limitations of satellite SARs for SWE estimation.

Research findings published so far revealed contradictory results on the effectiveness of extracting SWE from active microwave data.

Ulaby and Stiles (1980), using ground-based scatterometer data at 10.7 GHz, obtained a positive relationship between  $T_0$ , the radar backscatter, and SWE of a dry snow pack. They showed that the functional dependence of  $T_0$  on SWE takes the following form:

$$T_0 = C - D \exp(-\alpha_e \text{SWE}) \quad (1)$$

where  $\alpha_e$  is the effective two-way attenuation coefficient. It should be mentioned here that the snow pack used for the measurements was artificially piled and packed.

Matzler and Schanda (1984), using ground based scatterometer data at 10.4 GHz of a natural snow cover, did not observe any significant differences between the  $T_0$  values of a dry snow cover (20-64 cm SWE) and a snow-free surface. This behavior was explained by the large penetration depth (10 meters) at 10 GHz (Matzler and Schanda, 1984).

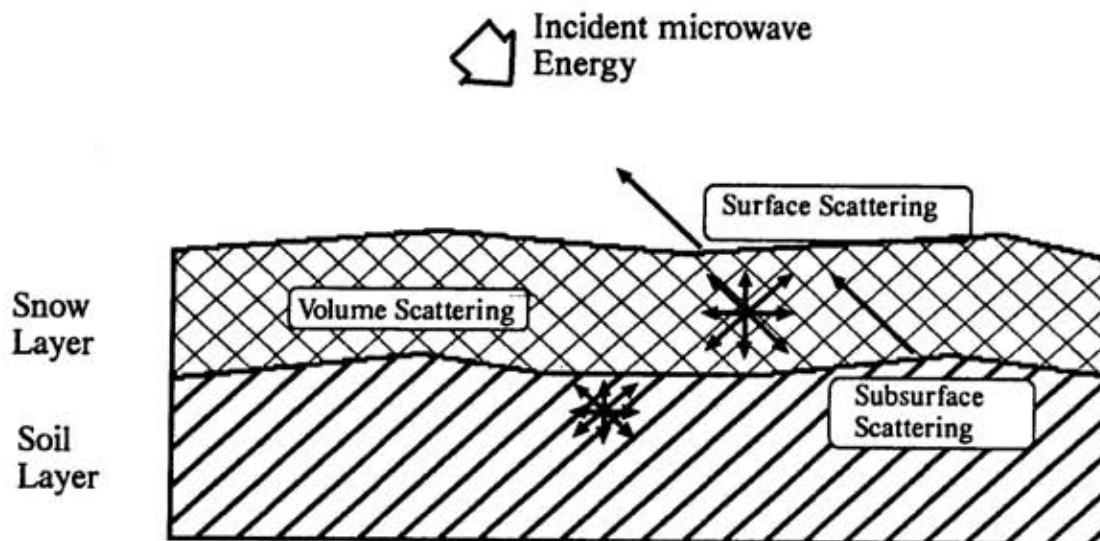
Leconte and Pultz (1990), using C-band airborne SAR data acquired over the course of a winter along a test line near Ottawa, Ont., Canada, obtained an inverse relationship between  $T_0$  and SWE. They suggested that soil surface roughness effects could account for the apparent contradictions between their results and results reported by Ulaby and Stiles (1980), with rough and smooth surfaces resulting respectively in negative and positive relationships. This hypothesis was further verified by Shi et al. (1990) using snow and soil backscatter models.

Finally Bernier (1991), using a multi-temporal airborne C-band SAR data set acquired during two winters near Sherbrooke, Que., Canada, suggested that the frozen/unfrozen state of the soil would play a fundamental role in the overall backscatter of a snow covered surface. A method was suggested to extract SWE from SAR data based on the thermal resistivity of the snow pack, as it influences the amount of unfrozen water in a frozen soil, and thus the dielectric constant at the snow/soil interface. The significant influence of a frozen/unfrozen soil layer on the surface backscattering coefficient was demonstrated by Wegmüller (1990).

The inconsistency in the research results obtained to date further suggests the need for more backscatter modelling studies. The usefulness of snow/soil backscatter models is that parameters susceptible to influence the microwave response of a snow covered surface can be identified; the sensitivity of the backscattering coefficient to changes of those parameters can be assessed; and the possibility that, and the optimal conditions under which, SWE and snow depth is extracted from radar data can be ascertained.

### The Model

The total backscatter of a snow covered surface is the sum of scattering at the snow/air interface, volume and multiple scattering within the snow pack, scattering at the snow/soil interface, and volumetric scattering in the upper soil layer, the latter two attenuated by the snow pack. Figure 1 illustrates the concept.



**Figure 1.** Backscatter mechanisms in snow pack.

When the snow is dry, the contribution from the air/snow interface is small because of the low dielectric contrast at the boundary ( $\epsilon' = 1$  and approx. 1.5-2 for air and dry snow, respectively), and therefore can usually be neglected. In addition, the soil penetration depth is usually small at the currently used SAR frequency range. Wegmüller (1990) reported penetration depths ranging from 0.2 to 1.5 cm and 1.2 to 6.5 cm in a wet unfrozen and in frozen soils, respectively, at frequencies ranging from 11 to 3.1 GHz. These small penetration depths suggest that volumetric scattering in the soil would be small. Results of experiments conducted at agricultural fields have shown penetration depths in the order of 5 to 10 centimeters at C-band (5.6 GHz), but that volumetric scattering in the first few centimeters of soil may be significant (Pultz, et al., 1990). As a first approximation, however, it will be assumed that the soil contribution to the total backscatter is from surface scattering only.

The backscatter model of a surface covered with dry snow therefore consists of a snow volumetric scattering component and a soil surface scattering component:



$$\sigma_t^p = \sigma_{sv}^p + t_p^2 l_f^2 \sigma_{gs}^p \quad (2)$$

where

- $\sigma_t$  = total backscattering coefficient
- $\sigma_{sv}$  = backscattering from the snow volume
- $t$  = power transmission coefficient
- $l_f^2$  =  $e^{-2\kappa SWE/\cos\theta t}$  = double pass loss factor
- $\kappa$  = mass extinction coefficient
- $\theta t$  = angle of transmission (related to incidence angle through Snell's law)
- $\sigma_{gs}$  = backscattering from the ground/snow interface
- $p$  = polarization (HH or VV)

This approach was proposed by Shi et al. (1990) and Ulaby et al. (1982).

The snow volumetric scattering was calculated using the random medium model proposed by Zuniga and Kong (1980) and Kong et al. (1980). Following this approach, a snow layer is described as a random medium with an average permittivity  $E_1$  and a random part characterized by a correlation function with a variance, a horizontal and a vertical correlation length. For simplicity, the snow pack in this study has been modeled as a one-layer random medium.

The small perturbation, physical optics, and geometric optics models (Ulaby, et al., 1982) were used to calculate the surface backscatter at the snow/ground interface at increasing levels of roughness. The models are all represented by the following equation:

$$\sigma_{gs}^p = |R_p|^2 S \quad (3)$$

where

- $R_p$  = Fresnel reflection coefficient at polarization  $p$
- $S$  = surface roughness parameter

As an example for a moderately rough surface, the physical optics approach (Ulaby et al., 1982) best describes the surface roughness parameter:

$$S = (kl\cos\theta)^2 \exp(-K_o) \quad (4)$$

where

$$K_o = 4k^2 s^2 \cos^2\theta$$

The parameters in equations (4) are:

- $k$  = wavenumber in free space
- $s$  = standard deviation of surface height
- $l$  = horizontal correlation length
- $\theta$  = incidence angle at the snow/ground interface

The Fresnel coefficients and the incidence angle at the snow/ground interface are a function of the dielectric constants of the snow and the ground. The dielectric constants are themselves function of snow and ground parameters. In particular the permittivity (real

part of the dielectric constant) of dry snow is a function of snow density. The soil permittivity is a function of soil liquid water content and soil texture. It is important to note here that soil permittivity increases with soil moisture. Therefore, an increase in soil moisture causes the backscattering coefficient to rise because the Fresnel coefficients increases.

Because there is a strong dependency between  $R_p$  and soil moisture, it is important to know whether the soil is frozen or not. A second model that simulates the thermal regime of a snow/soil system was coupled to the backscatter model previously introduced. This model calculates the temperature profile in the snow pack and in the soil and simulates the advance (or retreat) of a frozen soil layer.

The one dimensional heat conduction equation with phase change in a soil can be written as (Nakano and Brown, 1972):

$$\frac{\delta}{\delta z}(k_{so} \frac{\delta T_{so}}{\delta z}) - F \frac{\delta T_{so}}{\delta t} = 0 \quad (5)$$

where

$$\begin{aligned} z &= \text{vertical coordinate} \\ t &= \text{time} \\ T_{so} &= \text{soil temperature} \\ k_{so} &= \text{thermal conductivity of soil} \\ F &= C_{so} - L \frac{dG}{dT_{so}} = C_{so} + L \frac{dW_u}{dT_{so}} \\ C_{so} &= \text{volumetric heat capacity of soil} \\ L &= \text{latent heat of fusion} \\ G &= \text{ice content} \\ W_u &= \text{unfrozen water content} \end{aligned}$$

The term  $F$  in equation (5) takes into account the energy released/absorbed as water undergoes a phase change. The rate of change of  $W_u$  with  $T_{so}$  has been the subject of considerable research (Anderson and Tice, 1972; Black and Miller, 1990; Black, 1990). It was observed that, at soil temperatures below the freezing point, liquid water can be found, the amount increasing with decreasing pore size. The  $W_u - T_{so}$  model proposed by Anderson and Tice (1972) was used in this study.

Subsurface temperatures in a dry snow pack is often modeled as a simple problem of solid conduction without phase change, and neglecting radiation absorption which is effective only near the surface. The equation is:

$$\frac{\delta^2 T_{ds}}{\delta z^2} - \frac{1}{\alpha} \frac{\delta^2 T_{ds}}{\delta t} = 0 \quad (6)$$

where

$$\begin{aligned} \alpha &= k_{ds}/(\Gamma_{ds}C) \\ k_{ds} &= \text{thermal conductivity of dry snow ( a function of snow density)} \\ \Gamma_{ds} &= \text{snow density} \\ C &= \text{specific heat} \end{aligned}$$

The thermal model described by equations (5) and (6) was solved using a finite difference method with an implicit scheme for the time dimension. Boundary conditions were:

$$\begin{aligned} T_{\text{surf}} &= \text{air temperature} \\ T_{\text{bot}} &= \text{ground reference temperature} \end{aligned} \quad (7)$$

Radar backscatter was calculated by solving equations (2)-(7) in the following sequence: 1- the temperature profile in the snow pack and the soil is obtained by solving equations (5)-(7); 2- the unfrozen water content in the 0-5 cm upper soil layer is calculated using Anderson and Tice (1972) model; 3- the dielectric constant of the soil and Fresnel coefficients at the snow/soil interface are calculated, and; 4- equations (2) and (3) are solved for the radar backscatter. The effect of snow settlement on the backscattering coefficient at the snow/soil interface (through changes in the thermal regime of the snow/soil system) and on the volumetric scattering of the snow pack (through changes in the snow dielectric constant with snow density) was also taken into account by adding a snow compaction model (Riley, et al. 1969) to the above formulation.

The snow/soil backscatter/thermal model described by equations (2)-(7) was used in this study to investigate the following problems:

- how important is the soil type in calculating  $T_I$ ;
- what is the effect of a sudden drop in air temperature on  $T_I$ ;
- what is the effect of increasing SWE on  $T_I$ ;

Results however are valid within the limits imposed by the model formulation, that is:

- the snow pack is treated as a homogeneous one-layer system;
- the penetration depth in the soil is arbitrarily determined;
- the effect of vegetation is negligible.

## Results and Discussion

The simulations were carried out using a combination of snow depths and densities, soil types, and air temperature regimes. Typically, "equilibrium conditions" for a given snow depth and density and soil type were first created by running the model on a 30-day event, during which air temperature followed a daily sinusoidal pattern. As a first approximation, it was assumed that no precipitation has fallen during that 30-day event. Simulation runs were performed by assuming no and some degree of snow pack compaction. The ground and snow temperature regimes corresponding to the last day of the simulation were then used as initial conditions to "perturbation" runs, where the air temperature was modified. Table 1 illustrates the combination of parameters used in the "equilibrium" and the "perturbation runs".

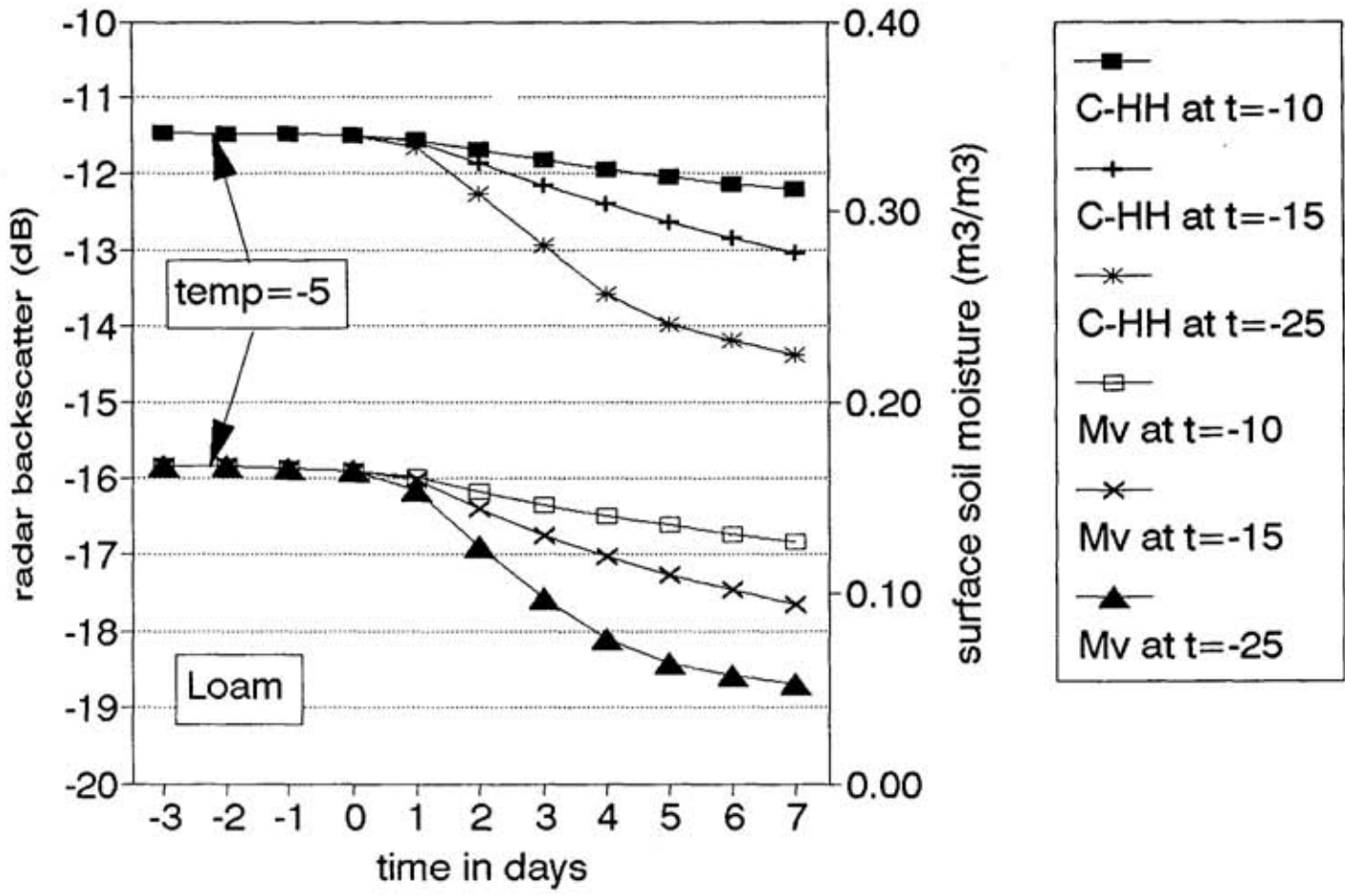
**Table 1.** Parameters used for the simulations.

Simulation run: 30 days for equilibrium runs 7 days for perturbation runs	
Frequency and polarization:	C-HH
Angle of incidence:	30 degrees
Surface scattering:	Physical optics model
Temperature regime: $-5\pm 4^{\circ}\text{C}$ , $-2\pm 1.5^{\circ}\text{C}$ for equilibrium runs $-10\pm 4^{\circ}\text{C}$ , $-15\pm 4^{\circ}\text{C}$ , $-20\pm 4^{\circ}\text{C}$ , $-25\pm 0^{\circ}\text{C}$ , for perturbation runs	
soil type :	Clay soil (6% sand, 47% clay, 47% silt) Till soil (15% sand, 45% clay, 40% silt) Loam soil (43% sand, 18% clay, 39% silt)
soil moisture:	20%, 30% gravimetric
Snow density:	$250 \text{ kg/m}^3$
Snow depth:	10, 50, 90 cm
Snow compaction:	no, yes

Figure 2 illustrates a typical example of an equilibrium / perturbation run for a 50-cm snow pack, with a snow density of  $250 \text{ kg/m}^3$ , over a loamy soil. The figure shows the simulated time series of radar backscatter  $T$ , as the system in equilibrium at an average  $T = -5^{\circ}\text{C}$  is perturbed at time = 0 by changing the average temperature  $-10$ ,  $-15$ , and  $-25^{\circ}\text{C}$ . As expected, a lowering of the air temperature reduced the unfrozen water content in the upper soil layer, which dropped from a value of 0.16 volumetric to approximately 0.05 after 7 days at  $T = -25^{\circ}\text{C}$ . This reduction in water content dropped the Fresnel coefficient at the snow/soil interface, with a corresponding decrease in the  $T$  value from  $-11.5$  to  $-14.5$  dB. The snow volumetric backscattering coefficient however has remained relatively constant with values ranging from  $-16$  to  $-17$  dB, making it a small contribution to the overall radar backscattering coefficient. This was expected because of the large penetration depths in dry snow packs at C-band, and also because snow is modeled as a homogeneous layer, with no ice lenses susceptible to generate multiple scattering in the pack.

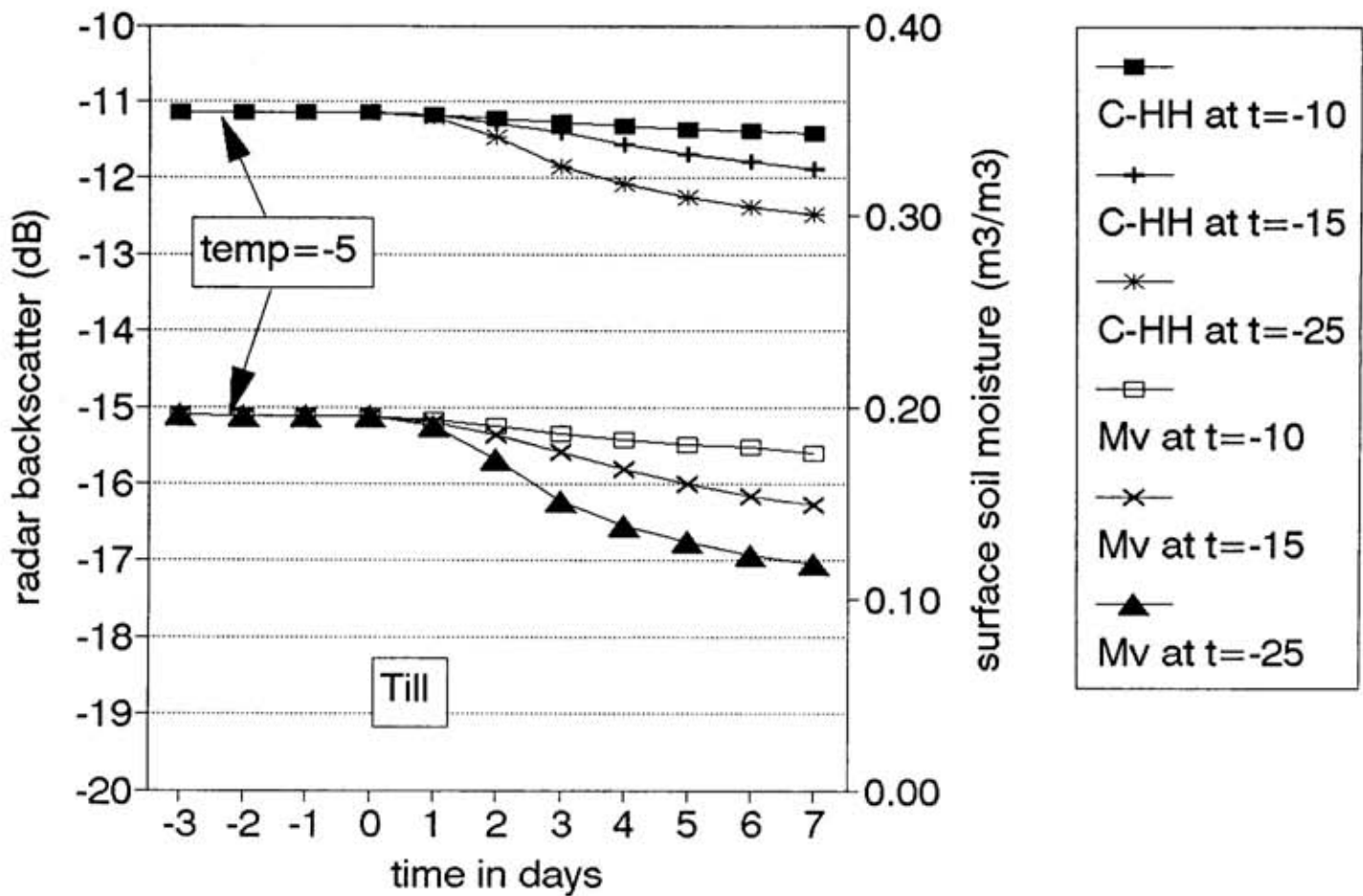
Figure 3 is similar to Figure 2, with the exception that the underlying soil is till, which is characterized by a higher clay content than the loam. While the same general pattern is observed here, that is a decrease in radar backscatter as the result of a temperature drop in the snow and soil, there are noticeable differences between the two figures. First, the radar backscattering coefficient for the snow covered till was  $-11$  dB at equilibrium, which is approximately 0.5 dB higher than for the snow covered loam. This is best explained by the differences in unfrozen water contents in the upper soil layer, which were approximately 0.19 volumetric for the till, as compared to 0.16 for the loam. This higher value was expected because a till contains a higher percent fine soil particles than a loam. The other noticeable difference is in the range of the system's response to a given temperature change. As shown in Figure 3, the  $T$  values ranged from  $-11$  to  $-12.5$  dB for the till, as compared to  $-11.5$  to  $-14.5$  dB for the loam. These variations in the dynamic range are

directly related to changes in unfrozen water content in the upper soil layer, which, as expected, were greater in the loam than in the till. Finally, simulation runs with the clay soil have revealed even higher T values (-11.2 dB) and a smaller dynamic range (0.8 dB) than for the till.



**Figure 2.** Influence of Temperature on Backscatter and surface soil moisture: 50-cm snow pack, loam soil.



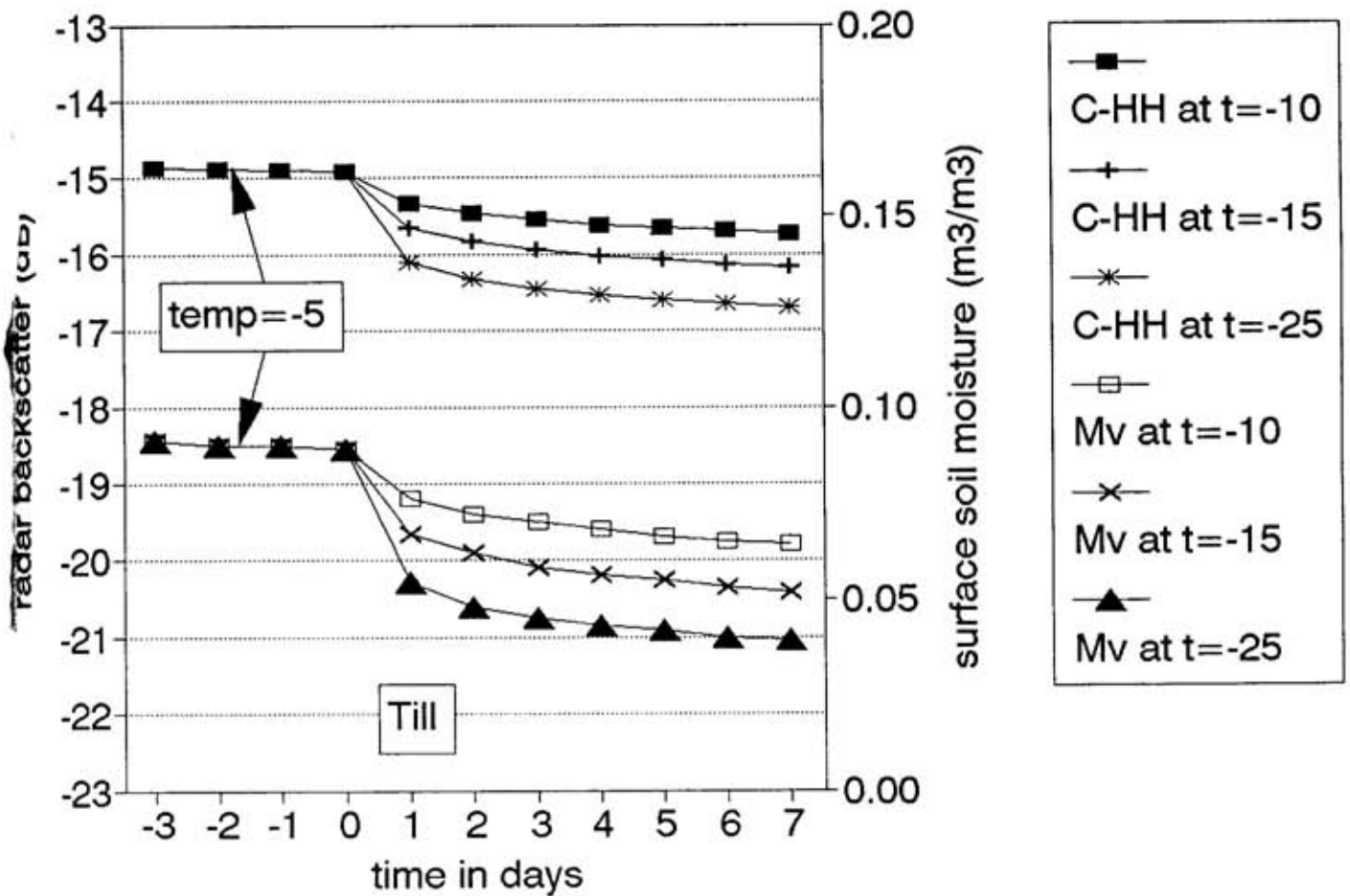


**Figure 3.** Influence of temperature on backscatter and surface soil moisture: 50-cm snow pack, till soil.

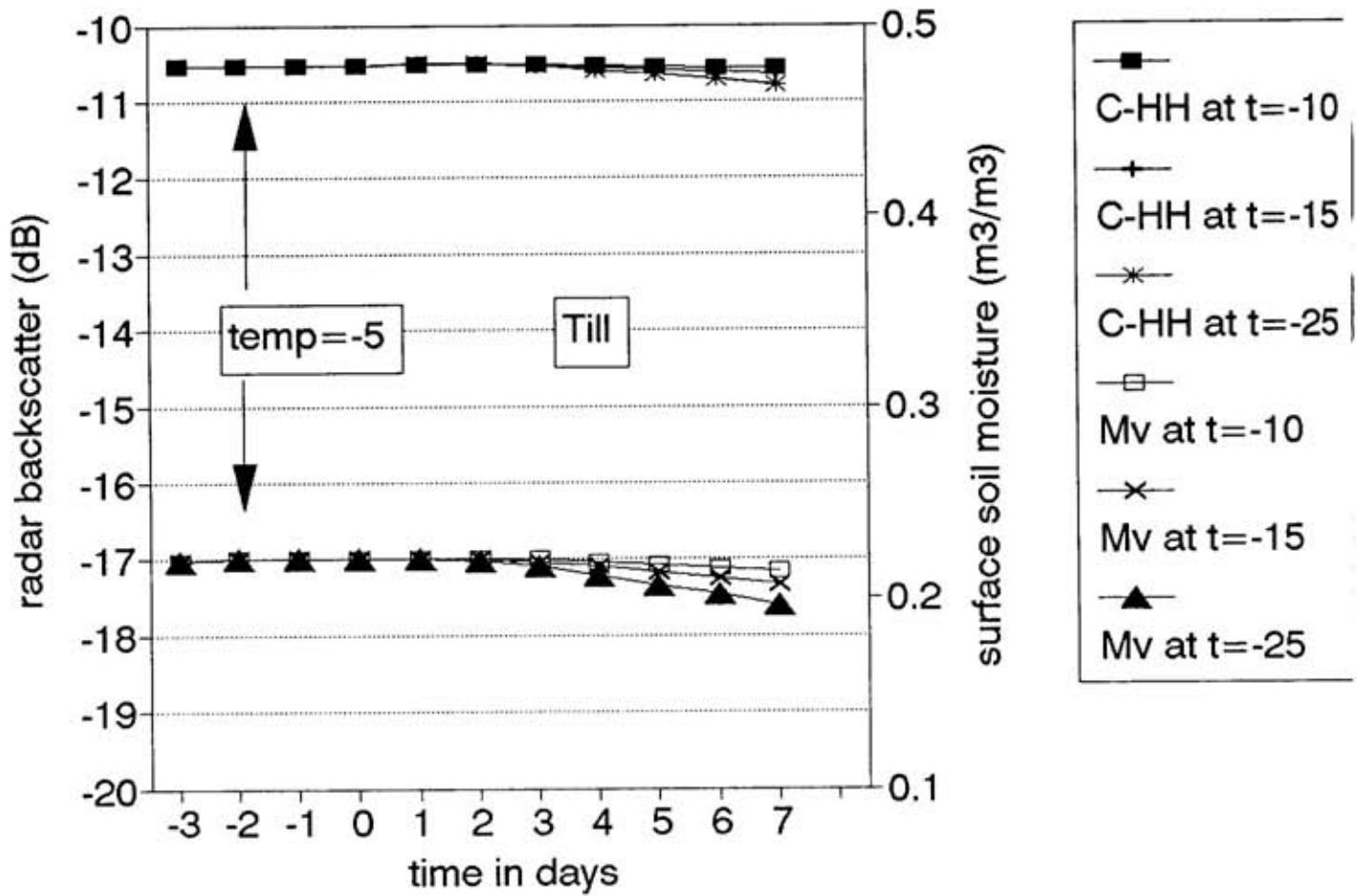
Because snow acts as an insulator, it is expected that varying amounts of snow will result in different snow/soil temperature regimes, which in turn will affect the radar backscattering coefficient through changes in unfrozen water content in the frozen soil. This is observed in Figures 4 and 5, for a 10- and a 90-cm snow pack on a till soil. The now familiar drop in the T values is again present in the 10-cm case shown in Figure 4, but is almost absent in the case of a 90-cm snow pack, as shown in Figure 5. By comparing the 10-, 50-, and 90-cm case shown in Figures 4, 3, and 5, respectively, one can observe that the radar backscatter coefficient at equilibrium for the snow covered till increased, and the dynamic range decreased, with increasing snow depths.

The strong dependency between the snow/soil thermal properties and the radar backscattering coefficient suggests that snow water equivalent/snow depth could, under certain conditions, be extracted from radar data. Figures 6 and 7 illustrate, for the clay, till, and loam soils used in the simulations, the variation of T with snow depth with a 250 kg/m<sup>3</sup> snow density, for both equilibrium (-5°C, Figure 6) and perturbed (-25°C, Figure 7) conditions. Generally, the radar backscattering coefficient appeared more sensitive to snow depth variations in the 0-50 cm depth range, and less sensitive in the 50-90 range. This was especially true of the coarser textured soils, where a temperature drop affected the soil liquid water content more than the fine textured soils. The reduced sensitivity with increasing snow depths is a direct consequence of the large thermal resistivity of the pack. Figures 6 and 7 also show the dependence of the T values with soil types, suggesting that, in order to extract snow depth or snow water equivalent from SAR data, a prior knowledge

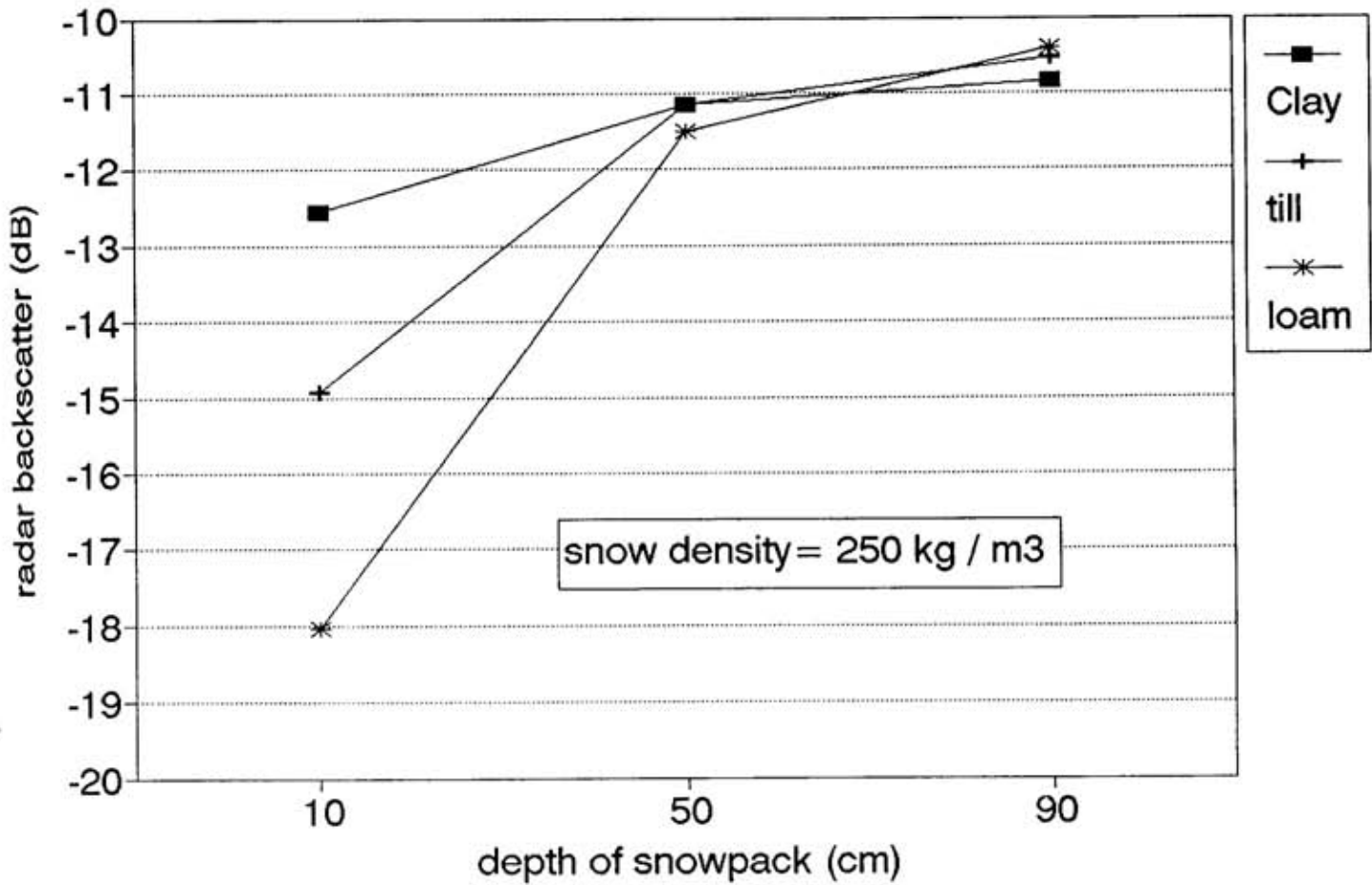
of the pedology of the study area would be required. Note also the dependence of the T - snow depth relationship on air temperature, suggesting that air temperature would also be an important factor to consider in extracting snow depth information. However, the slope of the T - snow depth relationship does not appear to vary appreciably with temperature, especially for coarser textured soils, implying that change detection techniques, coupled with a knowledge of snow depth at a few sites, may be adequate to obtain SAR-derived snow depth data.



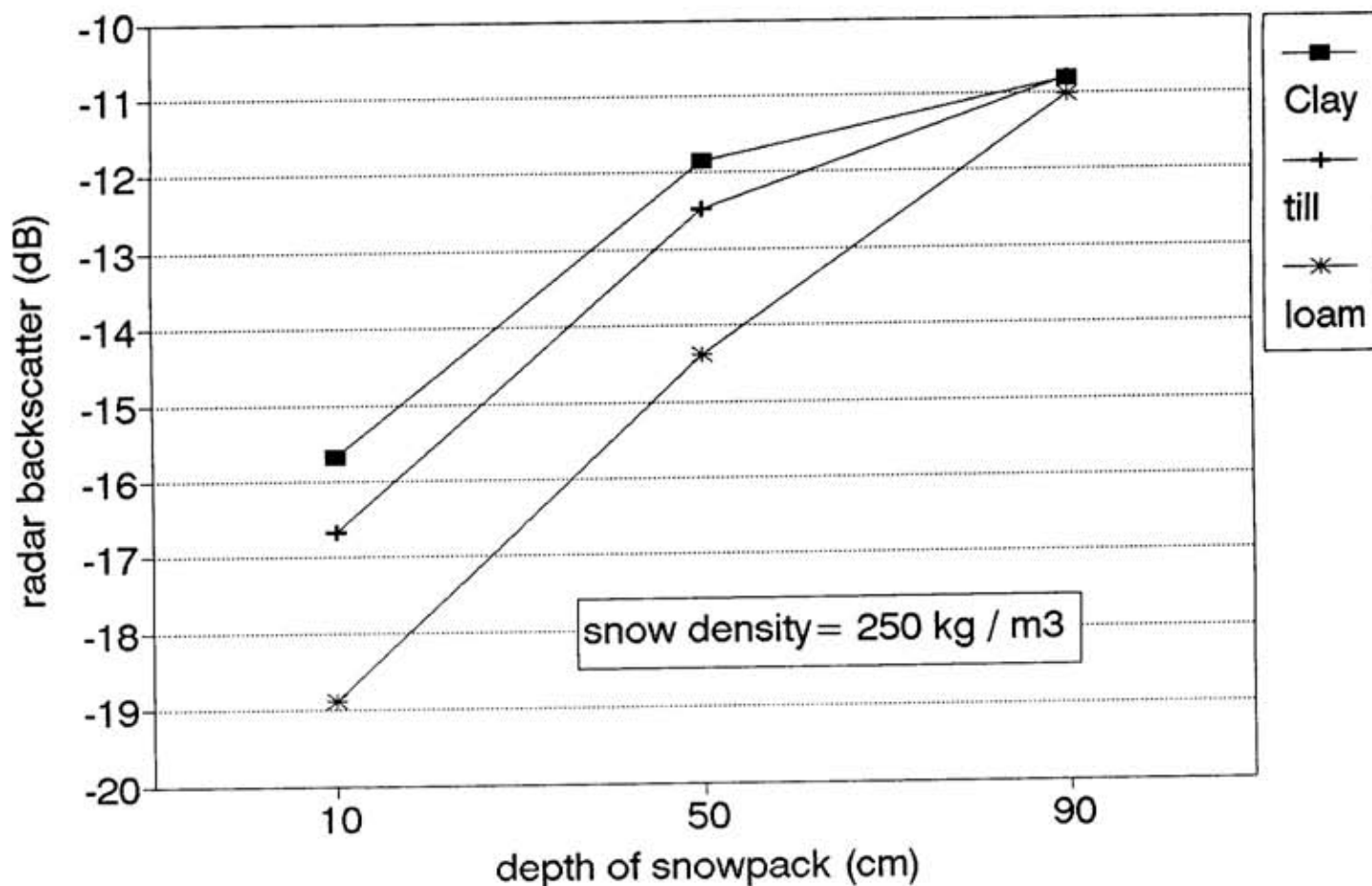
**Figure 4.** Influence of temperature on backscatter and surface soil moisture: 10-cm snow pack, till soil.



**Figure 5.** Influence of temperature on backscatter and surface soil moisture: 90-cm snow pack, till soil.



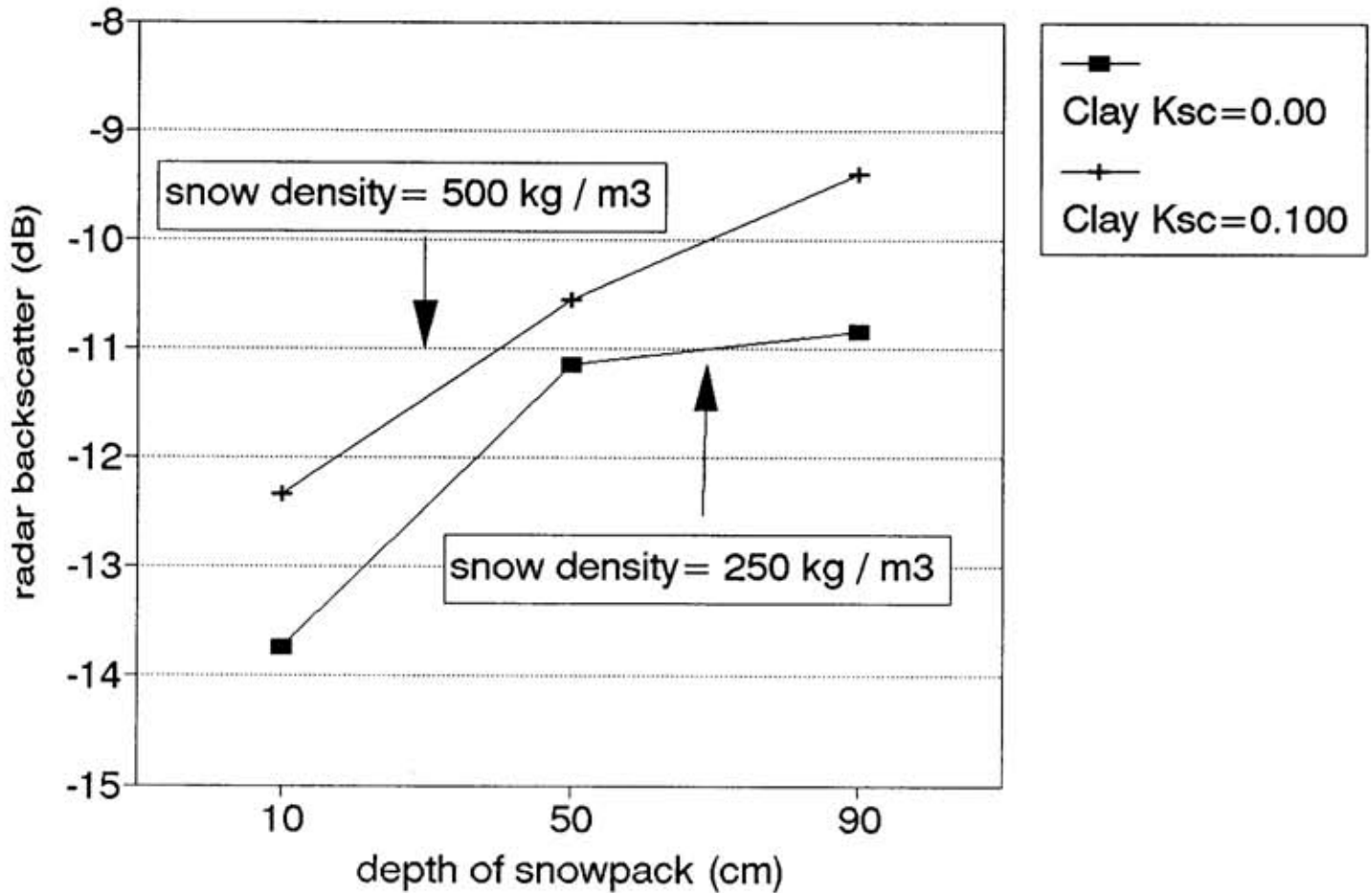
**Figure 6.** Influence of snow pack on backscatter for the loam, till, and clay soils: air temperature = -5°C and compaction coefficient=0.00.



**Figure 7.** Influence of snow pack on backscatter for the loam, till, and clay soils: air temperature =  $-25^{\circ}\text{C}$  and compaction coefficient = 0.00.

Many other factors have to be taken into account before any robust  $s$  - snow depth or SWE relationships can be derived. Target properties, such as soil moisture at the time of freeze-up, snow compaction and soil surface roughness, and sensor parameters, such as frequency and incidence angle, must be incorporated into such models. As an example, the simulations for the snow covered clay were repeated in the case where the snow pack was allowed to settle. The coefficient of settlement used in Riley's et al. (1969) model was 0.1. This resulted in an increase of the snow pack density from 250 to 500  $\text{kg}/\text{m}^3$  during a 30-day simulation period, which can be considered as an extreme case of snow compaction. Figure 8 compares the  $T$  - snow depth relationships for the snow compacted and no-compacted situations. The figure shows that introducing snow compaction resulted in an increase in radar backscatter and also in an increased sensitivity in deeper snow. Because the thermal resistivity of the snow pack decreased due to an increased snow density, the expected behavior was a decrease in radar backscatter, as the soil was more severely frozen. However, the increase in snow density also changed the microwave transmission angle in the snow pack, which became smaller. This in turn increased the surface roughness parameter (see equation (4)), offsetting the decrease resulting from a reduced soil moisture.





**Figure 8.** Influence of snow pack on backscatter for the clay soil: air temperature =  $-5^{\circ}\text{C}$  and compaction coefficient = 0.00 and 0.10.

### Conclusion

A snow/soil thermal/backscattering model has been developed to study the various mechanisms driving the radar backscattering of a snow covered terrain. More specifically, the influence of the unfrozen water content in a frozen soil was taken into account in the calculation of the surface backscattering at the soil/snow interface. It was demonstrated that, under certain conditions, snow depth and SWE can be extracted from radar data. This is because snow affects the amount of unfrozen water in a frozen soil, thereby affecting the soil dielectric constant and the radar backscattering coefficient at the snow/soil interface. More research however is needed to increase our knowledge of the microwave/snow/soil interactions before robust inversion models can be used to derive snow depth / SWE. This can best be achieved by a combination of modelling studies, where target and sensor parameters that drive the total backscatter are identified and interactions between the parameters are understood, and field experiments, where model results are compared with observations and inversion algorithms are developed and refined.

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# Lake Ice Conditions as a Cryospheric Indicator for Detecting Climate Variability in Canada

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## Abstract

The Atmospheric Environment Service (AES) database of surface observed lake ice condition dates is analyzed to determine the statistical properties of the lake ice data and to examine the degree of statistical relationship between composite lake ice condition dates and mean surface air temperatures. Also, composite lake ice condition time series are examined in order to detect evidence of regional climate variability, and to provide baseline climate/cryosphere relationships upon which impact related hypotheses might be developed. Eight lake ice regions are defined and strong statistical relationships are established between composite lake ice conditions and mean air temperatures from a station, or stations, in that region. Regional lake ice conditions appear to be useful indicators of temperature changes in that region during the spring and fall transition seasons. A 1°C change in any of the defined lake ice regions represents an average change of about 3 days in both freeze-up and break-up and a change of about 5 days in the duration of the ice and ice-free seasons. Trends during the past decade toward earlier and warmer spring seasons in western and central Canada are reflected in the earlier break-up of the lakes in most regions. No specific trends are seen in either air temperatures or regional freeze-up dates during the fall season in any part of the country. The trends towards shorter ice seasons, and longer ice-free seasons, in most regions of the country reflect the earlier spring break-up in response to warmer temperatures during the spring season.

## Introduction

The analysis of the temporal patterns of freeze-up and break-up of lake ice in Canada has both practical and theoretical value. The ability to accurately predict freeze-up and break-up dates would be of great benefit to marine and overland transportation operations which must take maximum advantage of lakes in their ice-free or ice covered states. The analysis of seasonal lake ice condition data could also provide early detection of a rise in global temperatures due to increased concentrations of carbon dioxide (CO<sub>2</sub>) and other radiatively active gases (RAGS). In addition, it would provide baseline climate/cryosphere relationships upon which climate change impact related hypotheses might be developed.

Freeze-up and break-up records from middle and high latitudes of North America could provide a good index of changing temperatures in the transition seasons. Trends toward later freeze-up and earlier break-up patterns could indicate warming due to increased atmospheric CO<sub>2</sub> concentrations (Tramoni et al., 1985). General circulation model experiments predict that higher latitudes will experience the greatest temperature increases (Manabe and Stouffer, 1980; National Academy of Sciences, 1982; Boer et al., 1990). This is mainly due to albedo feedback mechanisms and northern air mass stability. Carbon dioxide experiments with the Canadian Climate Centre general circulation model (Boer et al., 1990) indicate that the greatest warming would be north of 60°N in late fall-early winter with a secondary warming peak in the centre of the North American continent

in late winter-early spring. The freeze-up and break-up of lake ice in Canada occur at critical seasons for both the detection of CO<sub>2</sub>-related warming and the monitoring and assessment of its impacts.

The latitudinal belt between 50°N and 70°N is of vital importance to both present and future Canadian interests. It represents a potential transit route for increased shipping should industrial development occur in the Arctic. Oil and gas exploration, development and transportation will have to be conducted safely in order to protect both investments and the fragile environment. A warming due to increased CO<sub>2</sub> would likely result in a longer ice-free season on Canadian lakes and have significant impact on northern Canadian development. In addition, the ability to more accurately predict freeze-up and break-up dates would have added value in the planning of construction, logging, and mineral and oil exploration activities in the Arctic and subarctic.

The purpose of this study is to analyze the Canadian database of lake ice condition dates with sufficient length and completeness. The goal is to determine the statistical properties of the lake ice data, and to examine the degree of statistical relationship between composite lake ice condition dates and mean air temperatures. Also, composite lake ice condition time series are examined in order to detect any evidence of regional climatic change or variability, and to provide baseline climate/cryosphere relationships upon which climate change impact related hypotheses might be developed.

This represents the initial phase of an extended study examining changes in Canadian regional lake ice cover. It is viewed as an important part of overall cryospheric monitoring in response to global warming due to changing amounts of CO<sub>2</sub> and radiatively active gases in the atmosphere. The following phases of this study will use the surface observed ice condition data as ground truth for satellite monitoring of lake ice freeze-up and break-up and lake surface temperatures. Air-ice interaction models will be reviewed, developed and validated in both hindcast and forecast modes. Subsequently, an evaluation of the impact of observed or suspected trends in lake ice conditions on regional environmental conditions can be made.

## **Previous Studies**

Since the 1950's there has been a considerable amount of research into the study of lake ice growth and decay (for a review see Skinner, 1986). Recently, there has been a small amount of research dealing with freeze-up and break-up dates as indicators of climatic change. McFadden (1965) developed a method for estimating the relative depths of a group of lakes in Manitoba based on their freeze-up dates derived from aerial photography. Williams (1970) examined the limited number of available long-term North American break-up records for possible evidence of climatic variation. The break-up records were compared with long-term 10 year moving means of spring air temperature at Toronto with allowance made for urban effects. The ice cleared 10-15 days earlier in the 1950's than it did in the 1870's. This was accompanied by a significant increase in air temperature during that same period.

Da Silva (1984) analyzed lake ice freeze-up and break-up dates for 15 northern Canadian sites and correlated them with average monthly temperatures in order to detect climatic warming. He concluded that there were no distinct climatic trends in the data but there were reasonably good correlations between ice condition dates and local air temperatures. Da Silva (1985) developed multiple regression equations from average monthly air temperature, wind speed and cloud data for a number of Canadian stations to determine their importance on the freeze-up and break-up of lake ice. Monthly weather



values were weighted in order to develop a more realistic representation of conditions for the 30-day period prior to the freeze-up or break-up event. Stations were grouped into latitudinal belts and regression equations were calculated for each belt. The equations for each belt showed a high amount of variation.

Tramoni et al. (1985) have also analysed freeze-up and break-up records in middle and high latitudes of North America as possible indicators of climatic change. Regression analysis was used to test several temperature indices for a 50-day period, divided into 10 pentads, prior to a freeze-up or break-up event. This analysis indicated that there is no single significant pentad but pentad 6 (26-30 days) and pentad 8 (36-40 days) were selected more often and gave the highest correlation coefficients between freeze-up date and mean daily air temperature during the pentad. The time periods defined in this method, and in the Da Silva (1985) method, prior to the freeze-up or break-up event change from year-to-year with the changing ice condition dates. As a result, interpretational problems arise when dealing with the climate indices before the event.

Similar studies in Finland (Simojoki, 1959) report a ten day advance in break-up date in the 1940's and 1950's as compared with the 1860's. This is associated with about a 2°C spring temperature warming. In studying freeze-up dates for Lake Suwa, a mid-latitude lake in Japan, Tanaka and Yoshino (1982) show that a range of about 60 days in freeze-up corresponds with about a 4°C range in December to January mean temperatures. Tramoni et al., (1985) have indicated that for mid-latitude lakes an evenly distributed annual temperature rise of 1°C would delay freeze-up and advance break-up by about two to 15 days in each case. It appears that the sensitivity of lake ice to small temperature fluctuations is slightly less in higher latitudes, based on the results from a lake near Mould Bay, Northwest Territories, and greater at a mid-latitude lake such as Lake Suwa in Japan.

Palecki et al. (1985) summarized and expanded upon the previous work of Tramoni et al. (1985). Their data base is expanded to include 63 Finnish lakes from 60° to 70°N for the period 1959 to 1982. Freeze-up and break-up events were correlated with various temperature indices. Simple mean temperatures provided the best correlations. Temporal coverage was extended from 50 to 80 days prior to the lake ice event, to provide 16 rather than 10 pentads. Mean temperatures from each pentad were added consecutively from the first pentad back in time to yield cumulative mean temperatures for periods of varying length. Examination of individual pentads helps to identify key periods in the thermal history of the lake which in turn help to identify the ice-temperature relationship. This method also presents interpretational problems when dealing with the climate indices before the event. It examines the seasonal temperature trend instead of the response of the lake ice to seasonal temperatures because the temperature time period is non-static on a year-to-year basis (Palecki and Barry, 1986).

Palecki and Barry (1986) developed statistical relationships between lake ice freeze-up and break-up dates and air temperature means for year-to-year fixed time periods for the same 63 lakes in Finland. The interpretational problems of an interannual, non-static time period are avoided with this method. Significant correlations are evident for periods of up to five months prior to freeze-up. Regression coefficients are used to interpret changes in freeze up and break-up dates in terms of estimated changes in air temperature. A 1.1°C change in November air temperature would result in a five day change in freeze-up date in southern Finland.

Rousteenoja (1986) examined the relationship between spring weather conditions and the date of break-up of Lake Kallavesi ice in central Finland. Air temperature was found to satisfactorily explain the date of break-up, however, precipitation during the melting period also had some importance. Anomalies in radiation did not seem to have any significant

effect on the date of break-up. The use of break-up dates is strongly encouraged because of the close agreement between observed climatological temperature changes and the ones deduced on the basis of the ice conditions. Maslanik and Barry (1987) made a comparison of freeze-up and break-up dates of lakes in Canada and Finland using visible-wavelength images from the DMSP satellite and ground observations. Break-up dates from image interpretation are several days later than those from ground-based observations. Regional differences do exist in both the timing and variability of break-up. The use of visible satellite imagery in the fall is hampered by cloud cover. The use of satellite data analysis in regional climate variation studies appears viable with reasonable effort and cost.

Anderson (1987) evaluated the potential of using lake ice freeze-up and break-up dates as indicators of air temperature change. His methodology was applied to three northern Canadian lakes and results were encouraging. Significant correlations were found for both freeze-up and break-up with mean air temperatures over various time periods. Significant correlations were also found relating ice season duration to mean air temperatures for selected 12-month periods. Skinner (1988) applied this methodology to 12 other lake ice records in northern Canada and found similar results in some station records. Standard deviations of both the freeze-up and break-up dates varied from approximately seven to nine days. Regression equations developed for the individual records had small air temperature coefficients, meaning that it would be difficult to detect changes in mean air temperature of 1°C or 2°C based solely on the freeze-up or break-up dates.

Schindler et al. (1990) examined climatic and hydrologic records for the past 20 years for the Experimental Lakes Area of northwestern Ontario and found that air and lake temperatures had increased by about 2°C over the period. Also, the ice-free season duration has increased by about 20 days, due mainly to earlier break-up dates in the spring. Fall freeze-up dates were not observed to change significantly. Earlier spring break-up of the lakes in this area was seen as the result of two factors, the increased April-May air temperatures and reduced snow cover and warmer temperatures in March causing earlier snow melt and increased solar radiation absorption by the lake in early spring.

The main goal of this study is to examine the degree of statistical relationship between lake ice condition dates and mean air temperature. This is based on the assumption that air temperature is a major factor in the formation, growth and decay of lake ice. Other factors are of course recognized as important to these processes (Ragotzkie, 1978). Characteristics of lakes which determine their response to climate include fetch, mean and maximum depth, basin geometry and exposure to wind. Lakes also respond to climate factors besides air temperature. These include precipitation, shortwave, longwave and net radiation, wind, humidity and air stability above the surface of the lake. Ice thickness, snow depth and albedo are also important factors in the lake ice break-up process. Nevertheless, if significant relationships can be found between lake ice conditions and the air temperature regime, then the lake ice data base could prove beneficial in the study of climatic change.

The Canadian lake ice condition date database is maintained by the Atmospheric Environment Service (AES), Environment Canada (Allen, 1977). This database for lake, river and coastal sea ice sites contains information on the timing of freeze-up and break-up, maximum seasonal ice thickness, and the state of the ice surface with respect to the traffic it can support. (For a detailed description and critical analysis of the observed ice conditions see Anderson, 1987). The four standard observed ice condition dates are: first permanent ice (FPI), complete freeze over (CFO), first deterioration of ice (FDI), and water clear of ice (WCI). It was decided to focus on CFO and WCI dates because it was felt that these dates were not as subjective as the FPI and FDI dates and thus would be more accurately reported by observers. Also, time series of the CFO and WCI events for selected stations in

the Canadian Yukon and Northwest Territories in general showed less variability than those for the other two events (Anderson, 1987; Skinner, 1988).

Table 1 shows the 49 Canadian lake ice observation sites whose records were examined for this study. These observation sites are also shown in Figure 1. They were selected from the approximately 250 available lake ice sites in the database. Their records were manually scanned and determined to be of sufficient length and completeness for further analysis. Two stations, Ennadai Lake, NWT, and Nitchequon Lake, Quebec have been closed but had nearly complete long-term records. Each record is accompanied by the year when observations began and also by the last year of observations prior to this study. Record lengths vary from 18 years to 49 years. Fortunately, all lake ice observation stations are located at, or near, AES weather stations. As a result, reliable mean daily temperature data were readily available. Missing lake ice condition dates and missing daily temperature values were estimated by linear regression with corresponding records from a nearby station(s). In addition, all Julian dates for all lake ice records were adjusted by adding one day for each date past the last day in February for all leap years.

## Method

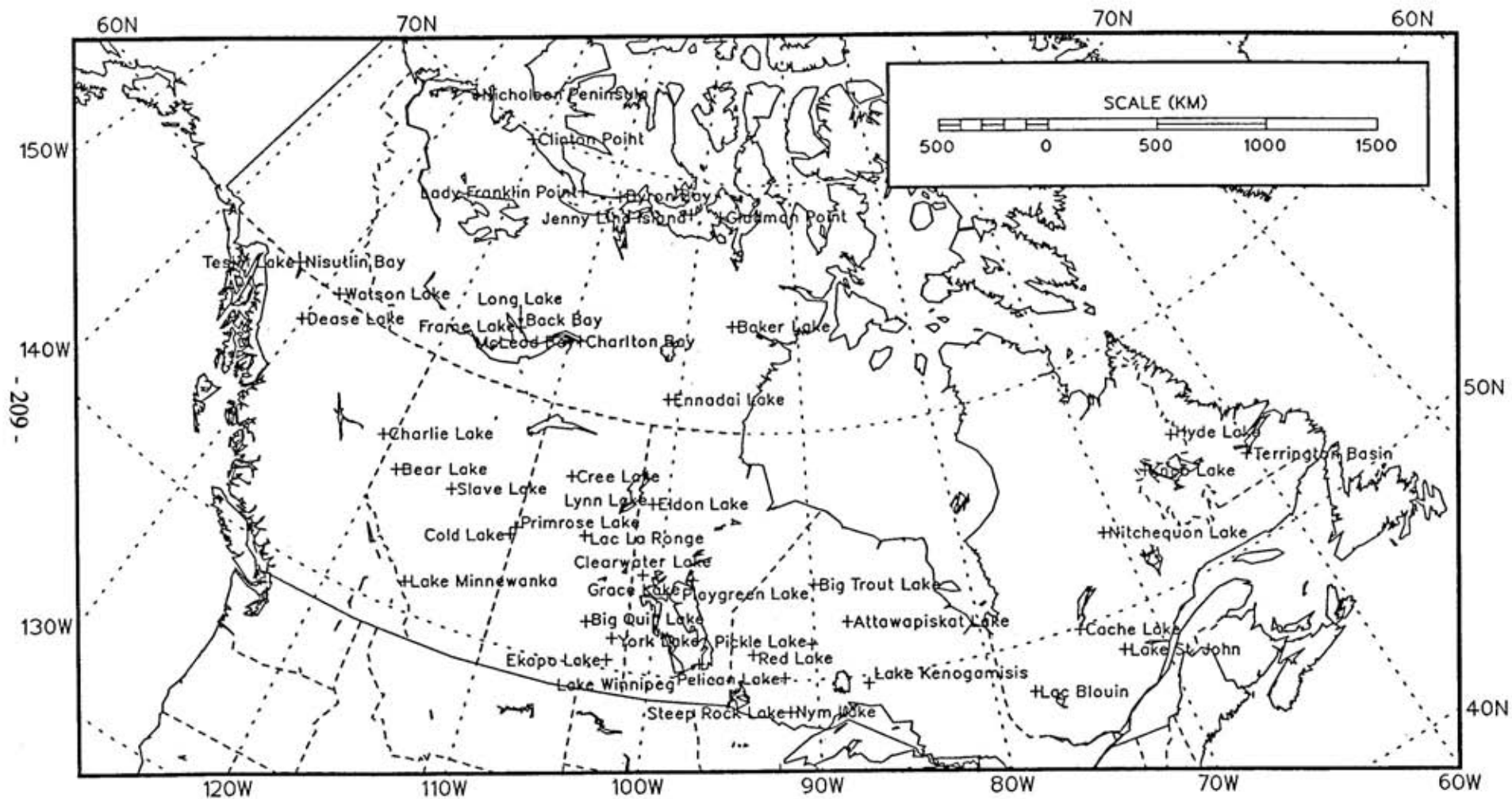
Simple statistics were derived for each of the 49 lake ice records as shown in Table 1. A number of these records, in the Northwest Territories, the Yukon and in northern Quebec, were previously analyzed in this manner (Anderson, 1987; Skinner, 1988). Table 2 shows the lake ice data record for Back Bay on Great Slave Lake near Yellowknife, Northwest Territories. This record represents a 'typical' series. It is 33 years long and had one missing CFO and WCI date. These dates were estimated by linear regression with other records on or near Great Slave Lake. The lengths of the ice season, and ice-free season, in days, were calculated from the freeze-up and break-up dates. The standard deviation for freeze-up dates is about 8.5 days while that for break-up is 7.5 days. Standard deviations for the ice season and ice-free season are 11.5 and 12 days, respectively. Regression equations developed for the individual CFO and WCI records (Skinner, 1988) had small air temperature coefficients, meaning that it would be difficult to detect changes in mean air temperature of 1°C or 2°C based solely on the freeze-up or break-up dates.

Palecki and Barry (1986) noted the locational difference of lake ice covers in Finland and analyzed a synthesis of northern Finnish lakes and southern Finnish lakes separately. These regions were further subdivided into coastal and inland zones. In this study, it was felt that the variability in the CFO and WCI dates might be reduced by combining similar lake ice records into regional lake ice indexes. Stronger relationships between ice condition dates and air temperature are possible when employing a regional composite of lake ice records rather than individual lake ice records (Palecki and Barry, 1986).

To accomplish this, correlation matrices for the CFO and WCI time series for 47 of the stations outlined in Table 1 were developed. The two stations that have closed were discarded as they were of no further use in the study. There was a 19 year period common to all lake ice records, ice years 1970-71 to 1988-89. It was felt that if the CFO and WCI time series for this period for two or more stations had a correlation value greater than a certain level then that group of stations were reacting to similar regional influences. Based upon previous studies, that similar regional influence was air temperature. Other factors such as lake size, shape and depth and man-made influences would be of secondary influence. An  $r$  value of 0.55 was selected. This value represents an  $\alpha=0.05$  at the 95% confidence level when  $N=19$ . Groups of stations that were intercorrelated above this level for both the CFO and WCI series were averaged to form regional indexes.



Figure 1. Canadian lake ice observation sites.



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Table 1. Individual Canadian lake ice observation sites.

<u>Lake</u>	<u>Latitude</u>	<u>Longitude</u>	<u>Ice Years</u>	<u>Years</u>
1. Baker Lake, NWT	64° 18'	96° 00'	1956-57 to 1988-89	33
2. Byron Bay, NWT	68° 45'	109° 04'	1966-67 to 1987-88	23
3. Ennadai Lake, NWT	61° 08'	100° 55'	1956-57 to 1979-80	33#
4. Charlton Bay, NWT	62° 43'	109° 06'	1956-57 to 1988-89	33*
5. McLeod Bay, NWT	62° 43'	109° 06'	1953-54 to 1988-89	36
6. Back Bay, NWT	62° 28'	114° 27'	1956-57 to 1988-89	33*
7. Frame Lake, NWT	62° 28'	114° 27'	1956-57 to 1988-89	33*
8. Long Lake, NWT	62° 28'	114° 27'	1955-56 to 1988-89	34*
9. Nicholson Peninsula, NWT	69° 54'	128° 58'	1962-63 to 1987-88	26
10. Clinton Point, NWT	69° 35'	120° 45'	1962-63 to 1987-88	26
11. Gladman Point, NWT	68° 40'	97° 48'	1962-63 to 1987-88	26
12. Jenny Lind Island, NWT	68° 39'	101° 04'	1964-65 to 1987-88	24
13. Lady Franklin Point, NWT	68° 30'	113° 13'	1963-64 to 1987-88	23
14. Nisutlin Bay, Yukon	60° 10'	132° 45'	1956-57 to 1988-89	33
15. Teslin Lake, Yukon	60° 10'	132° 45'	1949-50 to 1988-89	40*
16. Watson Lake, Yukon	60° 07'	128° 49'	1949-50 to 1988-89	40*
17. Dease Lake, BC	58° 25'	130° 00'	1956-57 to 1988-89	33*
18. Charlie Lake, BC	56° 14'	120° 44'	1958-59 to 1988-89	31
19. Lake Minnewanka, Alberta	51° 11'	115° 34'	1940-41 to 1988-89	49
20. Bear Lake, Alberta	55° 11'	118° 53'	1956-57 to 1988-89	33
21. Cold Lake, Alberta	54° 25'	110° 17'	1948-49 to 1988-89	41*
22. Primrose Lake, Alberta	54° 45'	110° 03'	1961-62 to 1988-89	28*
23. Slave Lake, Alberta	55° 21'	114° 59'	1953-54 to 1988-89	36*
24. Cree Lake, Saskatchewan	57° 21'	107° 08'	1970-71 to 1988-89	19*
25. Lac La Ronge, Saskatchewan	55° 08'	105° 20'	1967-68 to 1988-89	22*
26. Big Quill Lake, Saskatchewan	51° 46'	104° 12'	1956-57 to 1988-89	33
27. York Lake, Saskatchewan	51° 16'	102° 28'	1956-57 to 1988-89	33
28. Ekapo Lake, Saskatchewan	50° 23'	102° 35'	1956-57 to 1988-89	33*
29. Clearwater Lake, Manitoba	53° 58'	101° 06'	1956-57 to 1988-89	33*
30. Eldon Lake, Manitoba	56° 52'	101° 04'	1969-70 to 1988-89	20*
31. Grace Lake, Manitoba	53° 58'	101° 06'	1957-58 to 1988-89	32*
32. Lynn Lake, Manitoba	56° 52'	101° 04'	1969-70 to 1988-89	20*
33. Playgreen Lake, Manitoba	53° 59'	97° 50'	1970-71 to 1988-89	19*
34. Lake Winnipeg, Manitoba	50° 38'	97° 03'	1956-57 to 1988-89	33*
35. Red Lake, Ontario	51° 04'	93° 49'	1956-57 to 1988-89	33*
36. Big Trout Lake, Ontario	53° 50'	89° 52'	1952-53 to 1988-89	37*
37. Attawapiskat Lake, Ontario	52° 14'	87° 53'	1948-49 to 1988-89	41*
38. Lake Kenogamisis, Ontario	49° 41'	86° 57'	1967-68 to 1988-89	22*
39. Nym Lake, Ontario	48° 45'	91° 37'	1966-67 to 1988-89	23*
40. Pelican Lake, Ontario	50° 07'	91° 54'	1956-57 to 1988-89	33*
41. Pickle Lake, Ontario	51° 27'	90° 12'	1970-71 to 1988-89	19
42. Steep Rock Lake, Ontario	48° 45'	91° 37'	1966-67 to 1988-89	23*
43. Lac Blouin, Quebec	48° 03'	77° 47'	1961-62 to 1988-89	28
44. Cache Lake, Quebec	49° 54'	74° 18'	1971-72 to 1988-89	18*
45. Lake St. John, Quebec	48° 31'	72° 16'	1965-66 to 1988-89	24*
46. Knob Lake, Quebec	54° 48'	66° 49'	1955-56 to 1988-89	34*
47. Nitchequon Lake, Quebec	53° 12'	70° 54'	1951-52 to 1984-85	34#
48. Terrington Basin, NFLD	53° 19'	60° 25'	1951-52 to 1988-89	38
49. Hyde Lake, NFLD	55° 34'	64° 07'	1969-70 to 1988-89	20*

# station closed

\* station selected for further regional study based on cluster analysis



Table 2. Lake ice observation record for Back Bay,  
Great Slave Lake, NWT.

<u>Ice Year</u>	<u>CFO Date</u> <u>Julian</u>	<u>WCI Date</u> <u>Julian</u>	<u>Ice Season</u> <u>Days</u>	<u>Ice-Free Season</u> <u>Days</u>
1956 - 1957	291	162	236	142
1957 - 1958	304	145	206	172
1958 - 1959	317	165	213	126
1959 - 1960	291	156	230	140
1960 - 1961	296	154	223	139
1961 - 1962	293	161	233	149
1962 - 1963	310	146	201	176
1963 - 1964	322	157	200	149
1964 - 1965	306	158	217	148
1965 - 1966	306	149	208	147
1966 - 1967	296	159	228	149
1967 - 1968	308	161	218	147
1968 - 1969	308	152	209	157
1969 - 1970	309	153	209	135
1970 - 1971	288	144	221	159
1971 - 1972	303	154	216	156
1972 - 1973	310	142	197	165
1973 - 1974	307	147	205	136
1974 - 1975	283	141	223	159
1975 - 1976	300	137	202	172
1976 - 1977	309	145	201	163
1977 - 1978	308	149	206	149
1978 - 1979	298	160	227	143
1979 - 1980	303	139	201	174
1980 - 1981	313	144	196	149
1981 - 1982	293	158	230	143
1982 - 1983	301*	163	227	137
1983 - 1984	300	148	213	148
1984 - 1985	296	158	227	137
1985 - 1986	295	156	226	146
1986 - 1987	302	148	211	160
1987 - 1988	308	157	214	140
1988 - 1989	297	152	220	149
1989 - 1990	301			
<b>Mean:</b>	302.1	152.1	215.0	150.3
<b>Maximum:</b>	322	165	236	176
<b>Minimum:</b>	283	137	196	126
<b>Std. Dev.:</b>	8.4	7.5	11.5	12.3

\* estimated date

Table 3 shows the eight lake ice regions derived in this manner. Thirty of the 49 stations shown in Figure 1 were used. The Red Lake, Ontario record was used in two separate regional compositions. It is interesting to note that all of the stations located roughly along the 70th parallel in the western Northwest Territories were rejected from further analysis in this study. Other stations, such as Baker Lake and Lake Minnewanka, did not correlate well with any of the other stations. This does not mean that they do not possess valuable cryospheric information, just that they are distinct from the other stations in the study. Station records in some regions were well correlated with those in other regions such as the Saskatchewan-Manitoba, Manitoba-Ontario and the Northwest Ontario regions. For these regions, stations were grouped to maximize the length of at least one of the regional indexes, in this case the Manitoba-Ontario series at 33 years. Two lake ice regions located in eastern Canada, Southern Quebec and Labrador, had only two lake ice records each.

Table 4 shows the lake ice data record for the Great Slave Lake region, Northwest Territories. This record includes the Back Bay record presented in Table 2. This regional record retained its length at 33 years. The standard deviation for freeze-up dates is approximately 6 days, 2.5 days less than that for the individual record. The standard deviation for break-up is 7.5 days, the same as that for the individual lake ice record. Standard deviations for ice season and ice-free season are both about 1.5 days less than those for the individual lake ice record.

The next step in this analysis was to examine the degree of statistical relationship between the regional lake ice condition dates and mean air temperatures for a selected weather station in that region. Table 5 shows the air temperature stations used for this purpose. Palecki and Barry (1986) demonstrated that a single meteorological station could be successfully related to a large composited sample of lake ice records. All air temperature records were from single station records with the exception of that used for the Saskatchewan-Manitoba region where an average of daily mean temperatures from Lynn Lake A and The Pas A was used. Better correlations were found using the combined records than were found from the single station air temperature records. This was not surprising since this region contained the largest number and most widely distributed lake ice records.

A simple regression analysis technique was used to determine the relationships between mean air temperatures and the freeze-up and break-up dates, and ice season duration and ice-free season duration. Mean air temperatures for fixed calendar time periods prior to and/or during the lake ice event were used. Ten, or 11, day intervals of mean daily air temperatures were first calculated then regressed with the lake ice condition date time series. For example, during freeze-up, the month of October was divided into three intervals, October 1-10, 11-20, and 21-31 and during break-up, the month of April was divided into three intervals, April 1-10, 11-20, and 21-30. Several combinations of successive intervals, such as October 1-20, October 1-31 and October 1 to November 10, were subsequently regressed with the lake ice condition date time series.

Figures 2a to 2h show the regional time series of freeze-up (CFO), break-up (WCI), ice season duration and ice-free season duration for the eight lake ice regions. These figures also include the best-fit linear trend lines for each series. Tables 6a to 6h show the regression equations, correlations and standard errors for the eight lake ice regions for freeze-up (CFO), break-up (WCI), ice season duration and ice-free season duration for various periods. Each regional lake ice equation shows the correlation coefficient between mean air temperature and ice condition dates, the standard deviation of the slope (mean air

**Table 3. Lake ice regions defined on the basis of cluster analysis.**

<u>Region</u>	<u>Lakes</u>	<u>Ice Years</u>	<u>Years</u>
Great Slave Lake	Long Lake, NWT Frame Lake, NWT Back Bay (Great Slave Lake), NWT Charlton Bay (Great Slave Lake), NWT	1956-57 to 1988-89	33
Southern Yukon	Watson Lake, Yukon Teslin Lake, Yukon Dease Lake, BC	1956-57 to 1988-89	33
Alberta	Cold Lake, Alberta Slave Lake, Alberta Primrose Lake, Alberta	1961-62 to 1988-89	28
Saskatchewan - Manitoba	Cree Lake, Saskatchewan Lac La Ronge, Saskatchewan Lynn Lake, Manitoba Eldon Lake, Manitoba Grace Lake, Manitoba Clearwater Lake, Manitoba Little Playgreen Lake, Manitoba	1970-71 to 1988-89	19
Manitoba-Ontario	Big Trout Lake, Ontario Attawapiskat Lake, Ontario Red Lake, Ontario Lake Winnipeg, Manitoba Ekapo Lake, Saskatchewan	1956-57 to 1988-89	33
Northwest Ontario	Red Lake, Ontario Lake Kenogamisis, Ontario Pelican Lake, Ontario Steep Rock Lake, Ontario Nym Lake, Ontario	1967-68 to to 1988-89	22
Southern Quebec	Cache Lake, Quebec Lake St. John, Quebec	1971-72 to 1988-89	18
Labrador	Knob Lake, Quebec Hyde Lake, NFLD	1968-69 to 1988-89	23

**Table 4.** Mean lake ice observation record for the Great Slave Lake region as derived from Long Lake and Frame Lake, NWT and Back Bay and Charlton Bay, Great Slave Lake, NWT.

<u>Ice Year</u>	<u>CFO Date</u> <u>Julian</u>	<u>WCI Date</u> <u>Julian</u>	<u>Ice Season</u> <u>Days</u>	<u>Ice-Free Season</u> <u>Days</u>
1956 - 1957	297	163	232	142
1957 - 1958	305	152	212	156
1958 - 1959	307	168	226	124
1959 - 1960	292	159	232	140
1960 - 1961	298	157	224	137
1961 - 1962	294	163	234	144
1962 - 1963	307	149	208	166
1963 - 1964	315	157	207	148
1964 - 1965	305	159	220	145
1965 - 1966	304	152	213	147
1966 - 1967	299	164	230	146
1967 - 1968	310	161	216	152
1968 - 1969	312	160	213	148
1969 - 1970	307	156	214	147
1970 - 1971	302	146	209	159
1971 - 1972	305	160	220	137
1972 - 1973	297	144	212	164
1973 - 1974	308	152	209	139
1974 - 1975	291	147	221	160
1975 - 1976	306	137	196	171
1976 - 1977	307	151	209	161
1977 - 1978	311	153	207	152
1978 - 1979	305	164	224	138
1979 - 1980	301	143	207	173
1980 - 1981	315	150	200	150
1981 - 1982	299	166	232	136
1982 - 1983	302	166	230	134
1983 - 1984	300	150	215	149
1984 - 1985	299	162	228	138
1985 - 1986	299	158	224	148
1986 - 1987	305	152	212	152
1987 - 1988	304	160	221	143
1988 - 1989	303	156	218	151
1989 - 1990	307			
<b>Mean:</b>	303.3	155.4	217.2	148.1
<b>Maximum:</b>	315	168	234	173
<b>Minimum:</b>	291	137	196	124
<b>Std. Dev.:</b>	5.9	7.5	9.9	11.0

**Table 5.** Air temperature records used to determine statistical relationships with regional lake ice conditions.

<u>Lake Ice Region</u>	<u>Air Temperature Station</u>	<u>Ice Years</u>	<u>Years</u>
Great Slave Lake	Yellowknife Airport, NWT	1956-57 to 1988-89	33
Southern Yukon	Watson Lake Airport, Yukon	1956-57 to 1988-89	33
Alberta	Cold Lake Airport, Alberta	1961-62 to 1988-89	28
Saskatchewan - Manitoba	The Pas Airport, Manitoba Lynn Lake Airport, Manitoba	1970-71 to 1988-89	19
Manitoba-Ontario	Winnipeg Airport, Manitoba	1956-57 to 1988-89	33
Northwest Ontario	Sioux Lookout, Ontario	1967-68 to to 1988-89	22
Southern Quebec	Roberval Airport, Quebec	1971-72 to 1988-89	18
Labrador	Schefferville Airport, Quebec	1968-69 to 1988-89	23



Table 6a. Regression equations, correlations (r), standard errors (S<sub>e</sub>) and proposed changes in ice conditions due to the doubling of atmospheric CO<sub>2</sub> (CCC GCM) for the Great Slave Lake region. (N=33)

<u>Time Period</u>	<u>r</u>	<u>Regression Equation</u>	<u>S<sub>e</sub></u>	<u>2 x CO<sub>2</sub> (°C)</u>	<u>Change (Days)</u>
<b>CFO</b>					
October	0.89	y=307.3(±5.9)+2.33(±2.27)x	±2.7	+2.4	+5.6
11 Oct - 31 Oct	0.83	y=309.0(±5.9)+1.79(±2.74)x	±3.4		
21 Sep - 20 Oct	0.82	y=300.0(±5.9)+2.31(±2.08)x	±3.5		
01 Oct - 20 Oct	0.80	y=303.2(±5.9)+2.05(±2.29)x	±3.6		
<b>WCI</b>					
May	-0.82	y=166.5(±7.6)-2.33(±2.68)x	±4.4	+2.9	-8.0
21 Apr - 10 Jun	-0.88	y=168.3(±7.6)-2.75(±2.41)x	±3.7		
21 Apr - 31 May	-0.84	y=162.9(±7.6)-2.34(±2.73)x	±4.1		
01 May - 10 Jun	-0.86	y=173.0(±7.6)-2.82(±2.32)x	±3.9		
21 Apr - 20 May	-0.82	y=158.5(±7.6)-1.99(±3.10)x	±4.4		
<b>Ice Season</b>					
Jul - Jun	-0.60	y=192.6(±10.0)-4.74(±1.28)x	±8.1	+2.7	-14.3
Sep - Jun	-0.59	y=180.9(±10.0)-3.92(±1.51)x	±8.2		
Fall - Spring*	-0.89	y=225.1(±10.0)-5.29(±1.68)x	±4.7		
<b>Ice-Free Season</b>					
Apr - Oct	0.84	y= 95.4(±11.2)+11.31(±0.83)x	±6.2	+2.7	+14.0
May - Oct	0.80	y= 73.8(±11.2)+14.00(±0.64)x	±6.9		
Spring - Fall*	0.88	y=140.5(±11.2)+ 5.17(±1.89)x	±5.5		

\* Temperature average from the best correlated CFO and WCI series.

Table 6b. Regression equations, correlations (r), standard errors (S<sub>e</sub>) and proposed changes in ice conditions due to the doubling of atmospheric CO<sub>2</sub> (CCC GCM) for the southern Yukon region. (N=33)

<u>Time Period</u>	<u>r</u>	<u>Regression Equation</u>	<u>S<sub>e</sub></u>	<u>2 x CO<sub>2</sub> (°C)</u>	<u>Change (Days)</u>
<b>CFO</b>					
November	0.67	y=348.3(±6.7)+1.00(±4.53)x	±5.1	+3.8	+9.3
01 Nov - 10 Nov	0.75	y=343.6(±6.7)+0.95(±5.31)x	±4.5		
01 Sep - 30 Nov	0.81	y=356.5(±6.7)+2.45(±2.22)x	±4.0		
01 Nov - 10 Dec	0.76	y=355.7(±6.7)+1.33(±3.92)x	±4.3		
<b>WCI</b>					
May	-0.73	y=176.3(±6.0)-4.07(±1.08)x	±4.2	+2.9	-10.8
01 Apr - 20 May	-0.83	y=165.6(±6.0)-3.74(±1.33)x	±3.5		
01 Apr - 10 May	-0.73	y=158.1(±6.0)-2.72(±1.61)x	±4.2		
01 May - 20 May	-0.74	y=166.9(±6.0)-3.30(±1.36)x	±4.1		
<b>Ice Season</b>					
Oct - May	-0.74	y=134.3(±10.8)-4.23(±1.89)x	±7.4	+3.4	-13.0
Nov - May	-0.73	y=132.7(±10.8)-3.83(±2.06)x	±7.5		
Fall - Spring*	-0.82	y=156.8(±10.8)-3.81(±2.32)x	±6.3		
<b>Ice-Free Season</b>					
Apr - Nov	0.57	y=157.4(±8.9)+5.76(±0.88)x	±7.4	+3.4	+9.9
Spring - Fall*	0.65	y=203.1(±8.9)+2.91(±1.97)x	±6.9		

\* Temperature average from the best correlated CFO and WCI series.

Table 6c. Regression equations, correlations (r), standard errors (S<sub>e</sub>) and proposed changes in ice conditions due to the doubling of atmospheric CO<sub>2</sub> (CCC GCM) for the Alberta region. (N=28)

<u>Time Period</u>	<u>r</u>	<u>Regression Equation</u>	<u>S<sub>e</sub></u>	<u>2 x CO<sub>2</sub> (°C)</u>	<u>Change (Days)</u>
<b>CFO</b>					
November	0.74	y=348.9(±6.9)+1.35(±3.78)x	±4.8		
01 Nov - 10 Dec	0.84	y=353.9(±6.9)+1.67(±3.49)x	±3.8	+1.9	+3.2
11 Nov - 10 Dec	0.82	y=355.7(±6.9)+1.53(±3.72)x	±4.1		
21 Nov - 10 Dec	0.71	y=352.3(±6.9)+1.03(±4.79)x	±5.0		
<b>WCI</b>					
April	-0.79	y=144.5(±6.6)-2.16(±2.42)x	±4.1		
01 Apr - 10 Apr	-0.74	y=137.5(±6.6)-1.25(±3.91)x	±4.6		
01 Apr - 10 May	-0.84	y=148.7(±6.6)-2.48(±2.23)x	±3.7	+4.4	-10.9
01 Apr - 20 Apr	-0.77	y=140.5(±6.6)-1.79(±2.84)x	±4.3		
<b>Ice Season</b>					
Jul - Jun	-0.70	y=177.2(±10.6)-5.96(±1.25)x	±7.8		
Oct - May	-0.69	y=140.7(±10.6)-4.15(±1.76)x	±7.9		
Fall - Spring*	-0.85	y=154.7(±10.6)-3.95(±2.30)x	±5.7	+3.2	-12.6
<b>Ice-Free Season</b>					
Jan - Dec	0.71	y=194.4(±10.3)+6.42(±1.14)x	±7.4		
Apr - Nov	0.78	y=119.3(±10.3)+9.75(±0.82)x	±6.6		
Spring - Fall*	0.87	y=210.7(±10.3)+4.20(±2.13)x	±5.2	+3.2	+13.4

\* Temperature average from the best correlated CFO and WCI series.

Table 6d. Regression equations, correlations (r), standard errors (S<sub>e</sub>) and proposed changes in ice conditions due to the doubling of atmospheric CO<sub>2</sub> (CCC GCM) for the Saskatchewan-Manitoba region. (N=19)

<u>Time Period</u>	<u>r</u>	<u>Regression Equation</u>	<u>S<sub>e</sub></u>	<u>2 x CO<sub>2</sub> (°C)</u>	<u>Change (Days)</u>
<b>CFO</b>					
September	0.79	y=296.6(±5.1)+2.54(±1.59)x	±3.2		
Sep - Oct	0.83	y=301.5(±5.1)+3.23(±1.31)x	±2.9		
01 Sep - 20 Nov	0.87	y=312.5(±5.1)+2.72(±1.63)x	±2.6		
01 Sep - 10 Nov	0.91	y=306.6(±5.1)+3.18(±1.46)x	±2.1	+2.8	+8.9
10 Sep - 10 Nov	0.88	y=311.1(±5.1)+2.88(±1.55)x	±2.5		
<b>WCI</b>					
11 Apr - 20 May	-0.87	y=148.9(±8.1)-2.72(±2.59)x	±4.2	+2.1	-5.7
11 Apr - 10 May	-0.86	y=145.2(±8.1)-2.47(±2.84)x	±4.2		
21 Apr - 20 May	-0.81	y=151.9(±8.1)-2.58(±2.55)x	±4.9		
01 May - 20 May	-0.76	y=152.7(±8.1)-2.36(±2.63)x	±5.4		
<b>Ice Season</b>					
Fall - Spring*	-0.83	y=204.2(±8.6)-4.96(±1.44)x	±4.9	+2.5	-12.4
<b>Ice-Free Season</b>					
Spring - Fall*	0.87	y=157.8(±9.8)+5.78(±1.48)x	±5.1	+2.5	+14.5

\* Temperature average from the best correlated CFO and WCI series.

Table 6e. Regression equations, correlations (r), standard errors (S<sub>e</sub>) and proposed changes in ice conditions due to the doubling of atmospheric CO<sub>2</sub> (CCC GCM) for the Manitoba - Ontario region. (N=33)

<u>Time Period</u>	<u>r</u>	<u>Regression Equation</u>	<u>S<sub>e</sub></u>	<u>2 x CO<sub>2</sub></u> <u>(°C)</u>	<u>Change</u> <u>(Days)</u>
<b>CFO</b>					
01 Nov - 10 Nov	0.70	y=322.1(±5.9)+1.05(±3.96)x	±4.3		
01 Nov - 20 Nov	0.79	y=324.9(±5.9)+1.32(±3.54)x	±3.7	+3.0	+4.0
<b>WCI</b>					
April	-0.73	y=145.6(±6.6)-2.11(±2.29)x	±4.5		
May	-0.73	y=160.4(±6.6)-1.95(±2.47)x	±4.6		
01 Apr - 10 May	-0.79	y=149.0(±6.6)-2.26(±2.29)x	±4.1	+6.4	-14.5
<b>Ice Season</b>					
Fall - Spring*	-0.72	y=184.8(±8.8)-2.85(±2.23)x	±6.2	+4.7	-13.4
<b>Ice-Free Season</b>					
Jan - Dec	0.73	y=170.1(±9.0)+5.86(±1.10)x	±6.3		
Spring - Fall*	0.70	y=180.5(±9.0)+3.07(±2.25)x	±5.8	+4.7	+14.4

\* Temperature average from the best correlated CFO and WCI series.

Table 6f. Regression equations, correlations (r), standard errors (S<sub>e</sub>) and proposed changes in ice conditions due to the doubling of atmospheric CO<sub>2</sub> (CCC GCM) for the northwest Ontario region. (N=22)

<u>Time Period</u>	<u>r</u>	<u>Regression Equation</u>	<u>S<sub>e</sub></u>	<u>2 x CO<sub>2</sub></u> <u>(°C)</u>	<u>Change</u> <u>(Days)</u>
<b>CFO</b>					
November	0.70	y=330.4(±4.7)+1.37(±2.43)x	±3.5		
01 Nov - 10 Nov	0.67	y=324.2(±4.7)+0.82(±3.86)x	±3.6		
01 Nov - 20 Nov	0.71	y=326.9(±4.7)+1.16(±2.89)x	±3.4	+2.6	+3.0
<b>WCI</b>					
11 Mar - 20 Apr	-0.83	y=120.1(±6.6)-2.23(±2.48)x	±3.7	+2.7	-6.0
11 Mar - 10 Apr	-0.76	y=117.2(±6.6)-1.88(±2.66)x	±4.4		
21 Mar - 20 Apr	-0.82	y=124.0(±6.6)-1.86(±2.91)x	±3.9		
01 Apr - 20 Apr	-0.81	y=127.3(±6.6)-1.71(±3.11)x	±4.0		
<b>Ice Season</b>					
Fall - Spring*	-0.59	y=159.1(±7.2)-2.96(±1.42)x	±6.0	+2.7	-8.0
<b>Ice-Free Season</b>					
Jan - Dec	0.68	y=192.4(±6.4)+3.94(±1.09)x	±4.8		
Spring - Fall*	0.65	y=205.0(±6.4)+2.61(±1.59)x	±5.0	+2.7	+7.0

\* Temperature average from the best correlated CFO and WCI series.

Table 6g. Regression equations, correlations (r), standard errors (S<sub>e</sub>) and proposed changes in ice conditions due to the doubling of atmospheric CO<sub>2</sub> (CCC GCM) for the southern Quebec region. (N=18)

<u>Time Period</u>	<u>r</u>	<u>Regression Equation</u>	<u>S<sub>e</sub></u>	<u>2 x CO<sub>2</sub> (°C)</u>	<u>Change (Days)</u>
<b>CFO</b>					
November	0.65	y=340.6(±6.5)+2.78(±1.50)x	±5.1	+1.3	+3.6
<b>WCI</b>					
April	-0.83	y=139.3(±7.5)-2.97(±2.08)x	±4.3		
21 Mar - 31 Mar	-0.85	y=128.8(±7.5)-1.38(±4.57)x	±4.1		
21 Mar - 10 May	-0.96	y=139.6(±7.5)-3.30(±2.16)x	±2.2	+1.8	-5.9
21 Mar - 30 Apr	-0.90	y=135.2(±7.5)-2.68(±2.51)x	±3.3		
21 Mar - 20 Apr	-0.88	y=131.6(±7.5)-2.20(±2.99)x	±3.6		
21 Mar - 10 Apr	-0.84	y=129.1(±7.5)-1.77(±3.56)x	±4.1		
<b>Ice Season</b>					
July - June	-0.55	y=191.3(±8.6)-11.8(±0.55)x	±5.9		
Fall - Spring*	-0.63	y=163.0(±8.6)-6.67(±1.03)x	±5.4	+1.6	-10.7
<b>Ice-Free Season</b>					
Jan - Dec	0.66	y=183.3(±10.7)+7.80(±0.91)x	±8.3		
Spring - Fall*	0.84	y=202.8(±10.8)+7.32(±1.23)x	±6.1	+1.6	+11.7

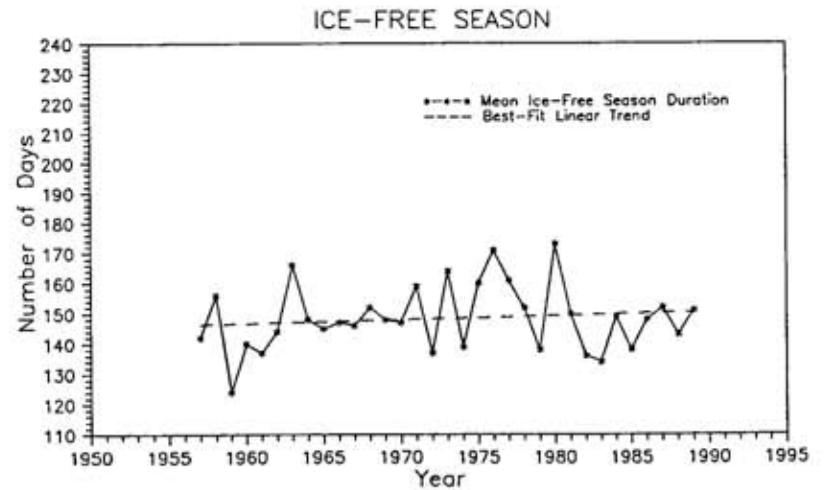
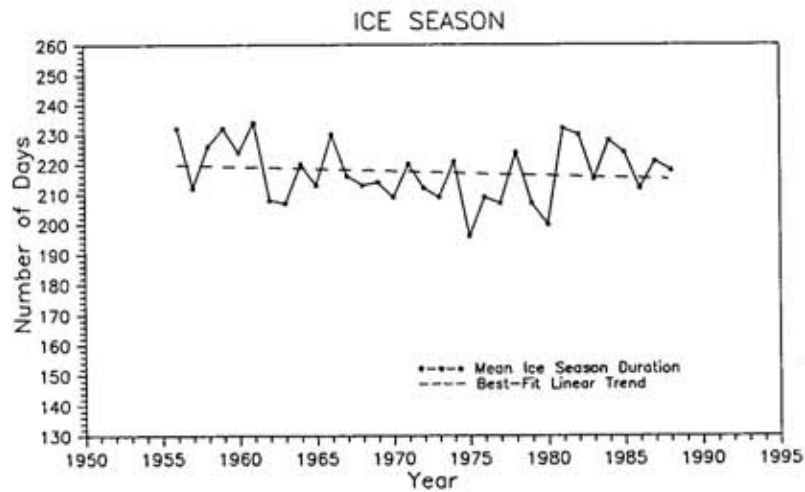
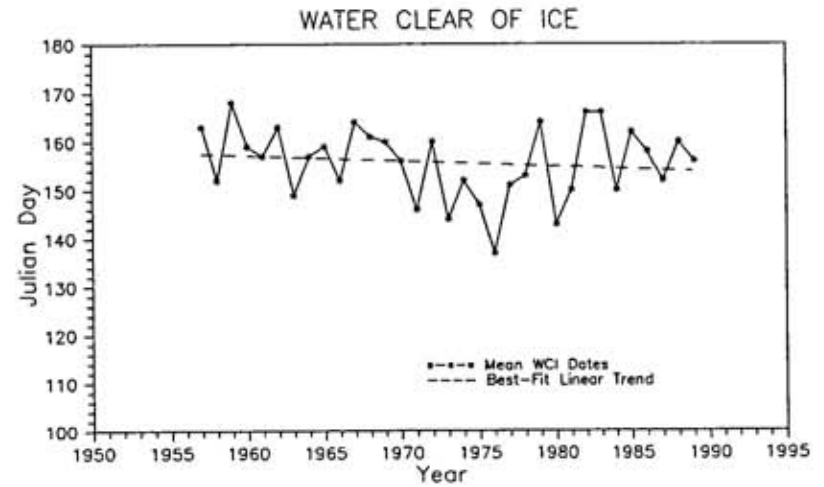
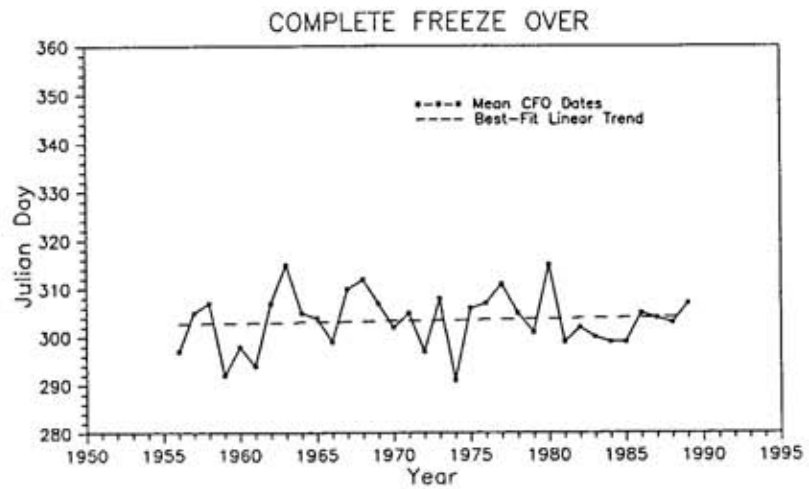
\* Temperature average from the best correlated CFO and WCI series.

Table 6h. Regression equations, correlations (r), standard errors (S<sub>e</sub>) and proposed changes in ice conditions due to the doubling of atmospheric CO<sub>2</sub> (CCC GCM) for the Labrador region. (N=23)

<u>Time Period</u>	<u>r</u>	<u>Regression Equation</u>	<u>S<sub>e</sub></u>	<u>2 x CO<sub>2</sub> (°C)</u>	<u>Change (Days)</u>
<b>CFO</b>					
21 Sep - 10 Nov	0.73	y=311.8(±9.0)+4.36(±1.50)x	±6.3		
21 Sep - 31 Oct	0.74	y=311.9(±9.0)+4.38(±1.51)x	±6.2	+1.9	+8.3
21 Oct - 10 Nov	0.71	y=319.5(±9.0)+3.02(±2.09)x	±6.5		
<b>WCI</b>					
11 Apr - 31 May	-0.81	y=158.5(±8.6)-3.25(±2.14)x	±5.2	+2.7	-8.8
11 Apr - 20 May	-0.79	y=155.7(±8.6)-2.90(±2.34)x	±5.4		
11 Apr - 10 May	-0.77	y=152.8(±8.6)-2.55(±2.58)x	±5.7		
11 Apr - 30 Apr	-0.73	y=152.9(±8.6)-1.88(±3.33)x	±6.1		
<b>Ice Season</b>					
Fall - Spring*	-0.53	y=217.1(±12.9)-5.05(±1.36)x	±11.2	+2.3	-11.6
<b>Ice-Free Season</b>					
Spring - Fall*	0.66	y=148.9(±12.5)+5.68(±1.45)x	±9.6	+2.3	+13.1

\* Temperature average from the best correlated CFO and WCI series.

Figure 2a. Lake ice condition time series for the Great Slave Lake region.





**Figure 2b.** Lake ice condition time series for the Southern Yukon region.

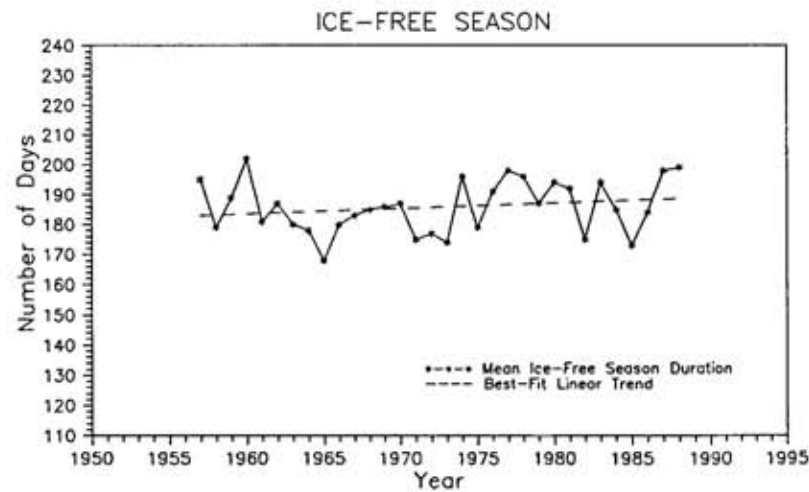
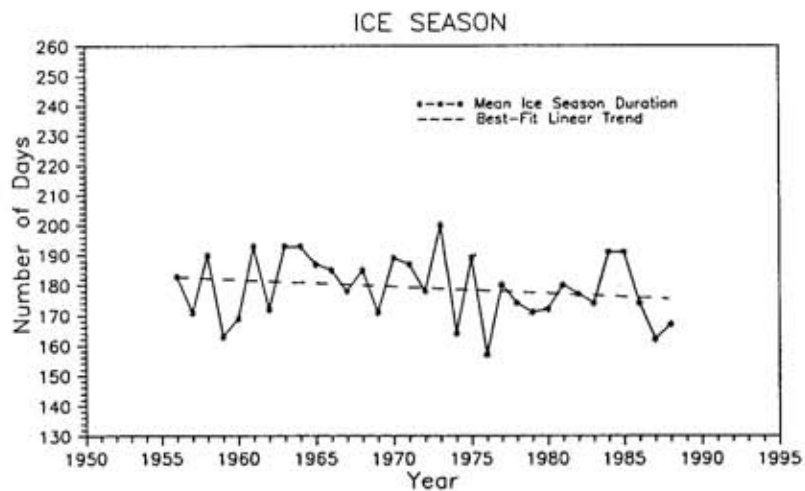
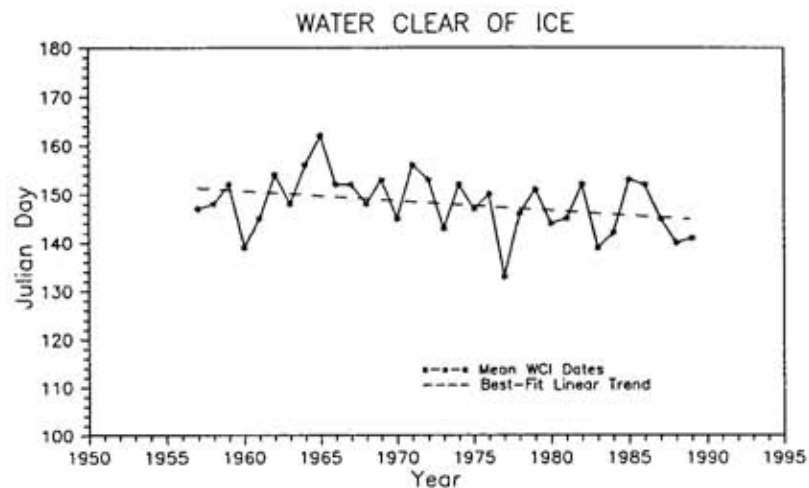


Figure 2c. Lake ice condition time series for the Alberta region.

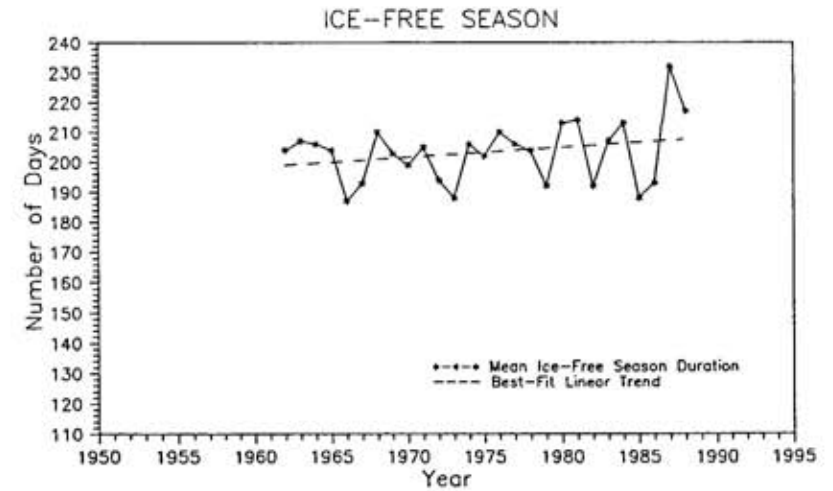
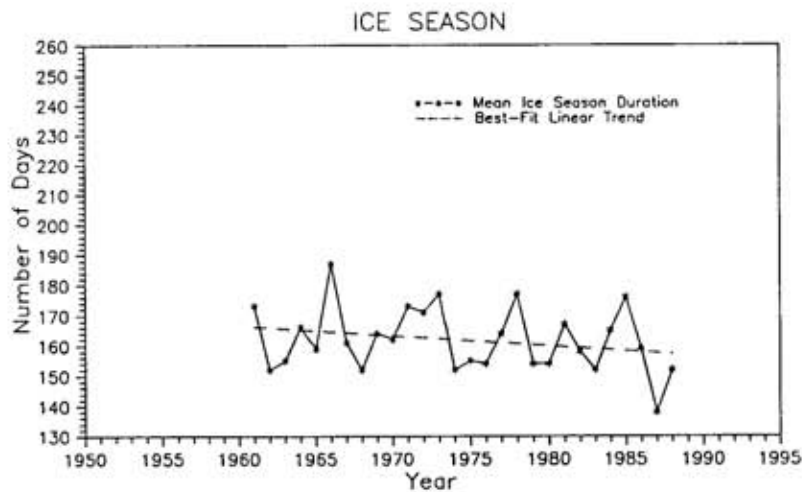
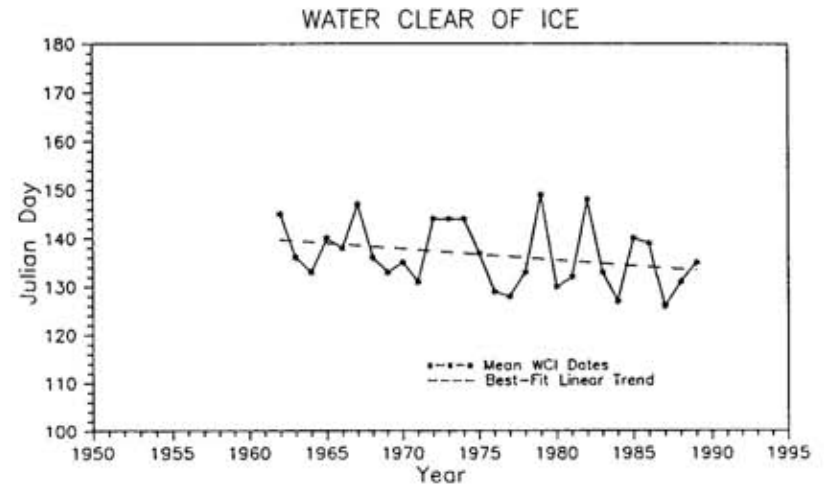


Figure 2d. Lake ice condition time series for the Saskatchewan-Manitoba region.

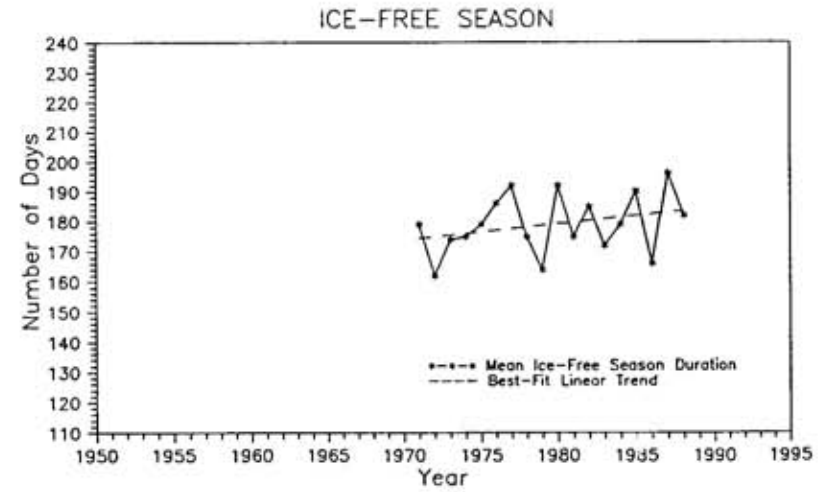
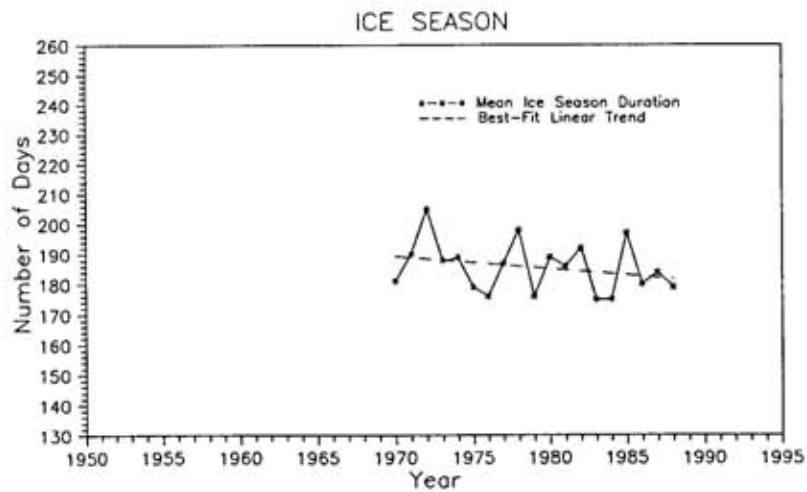
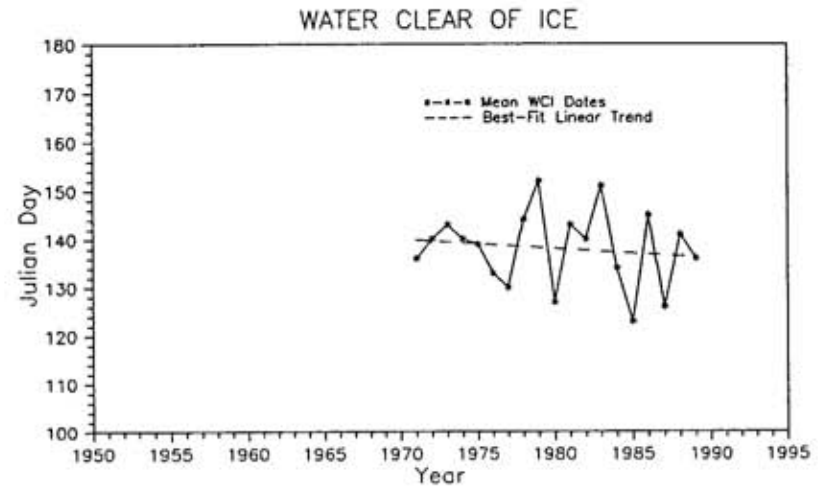


Figure 2e. Lake ice condition time series for the Manitoba-Ontario region.

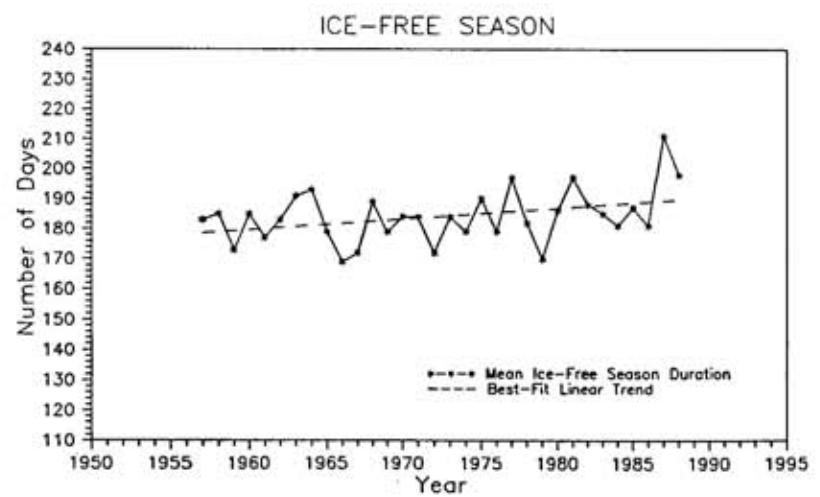
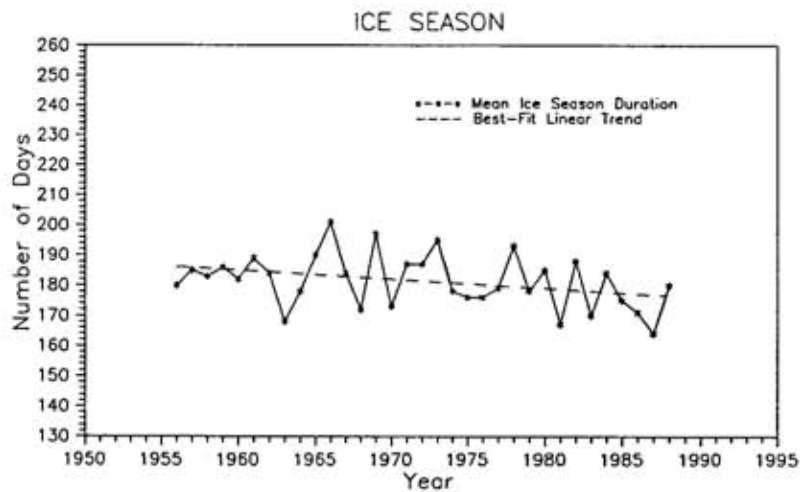
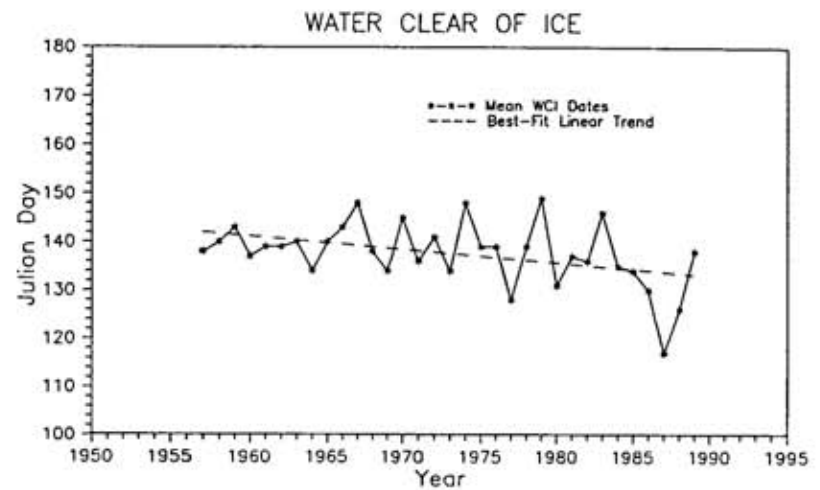
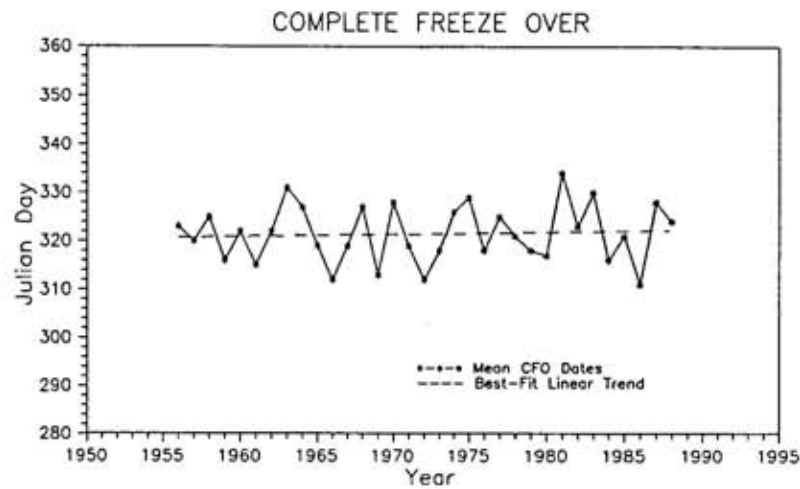


Figure 2f. Lake ice condition time series for the Northwest Ontario region.

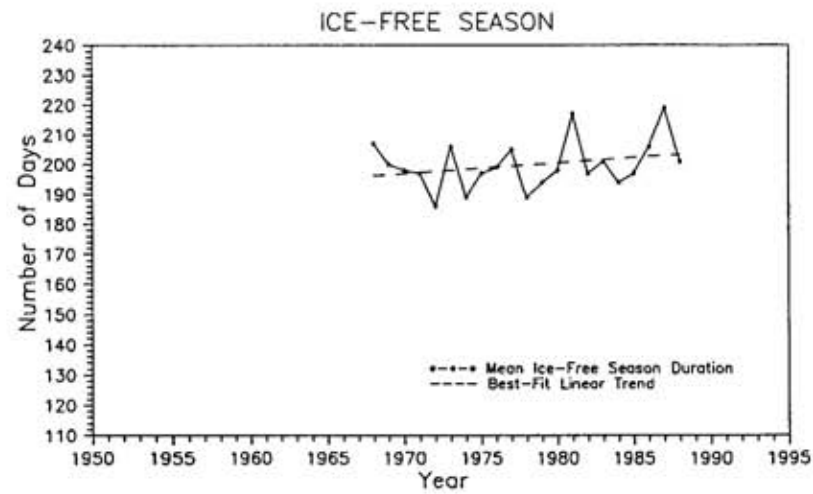
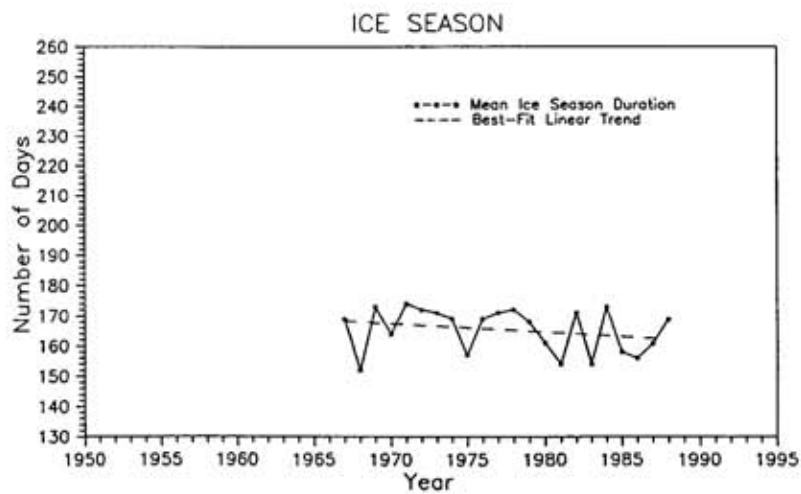
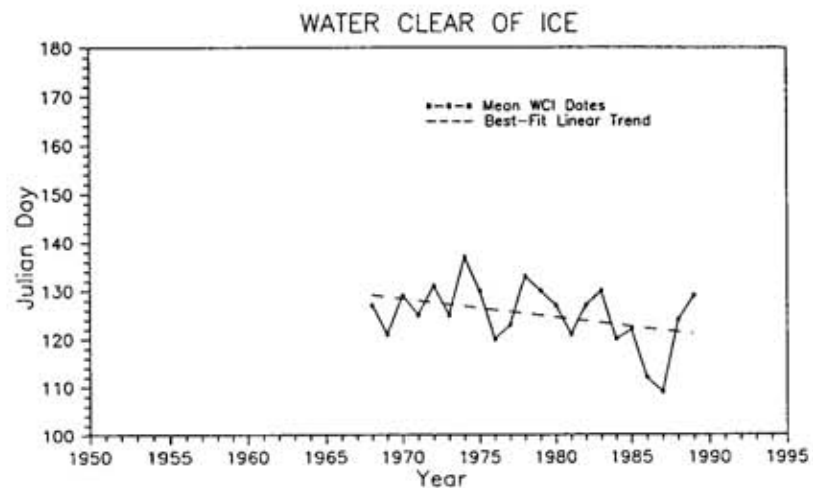




Figure 2g. Lake ice condition time series for the Southern Québec region.

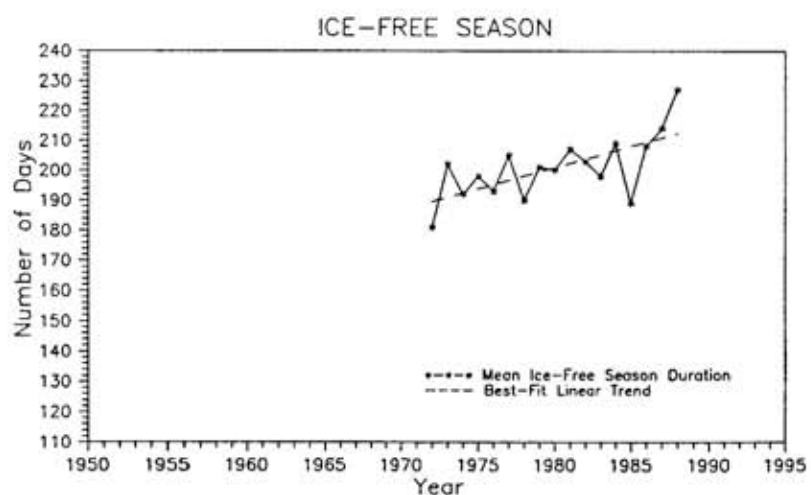
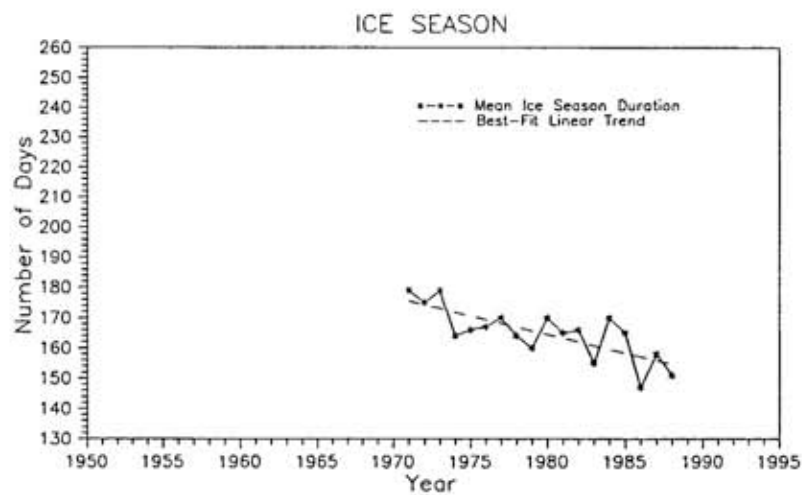
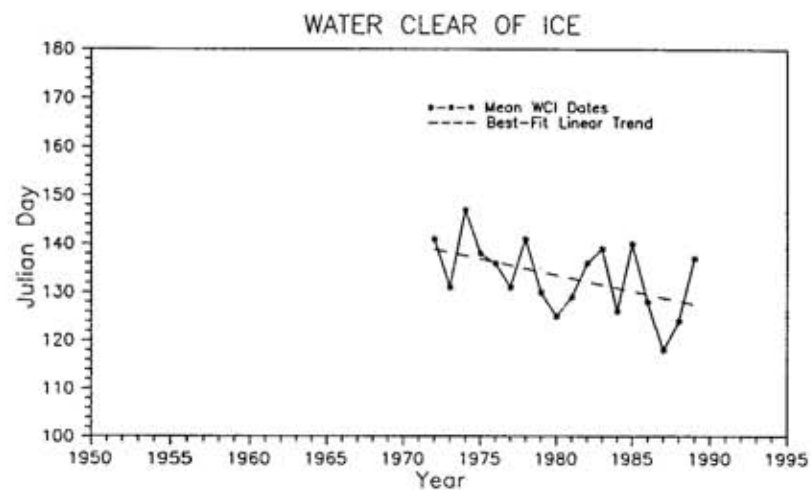
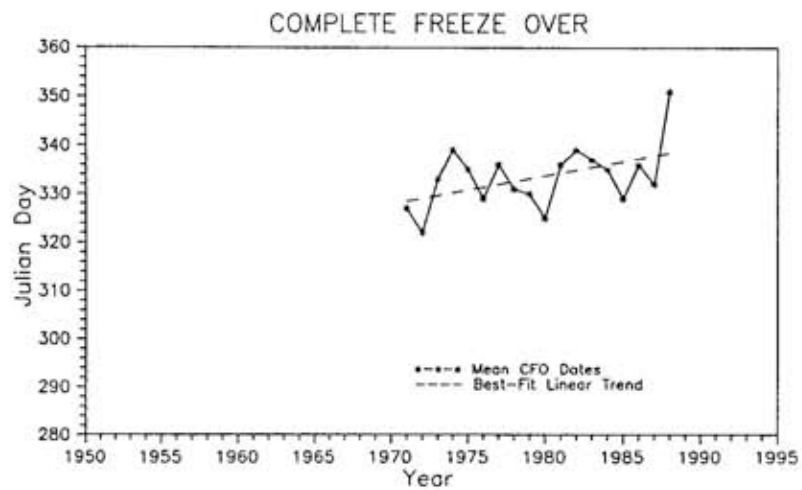
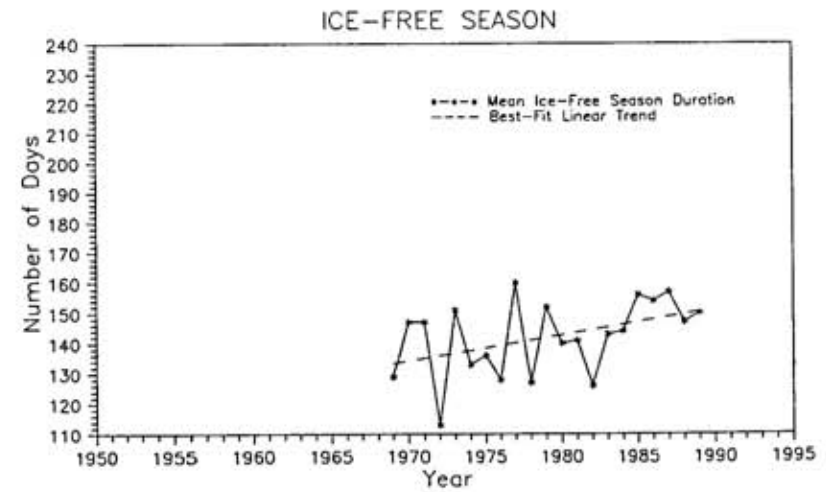
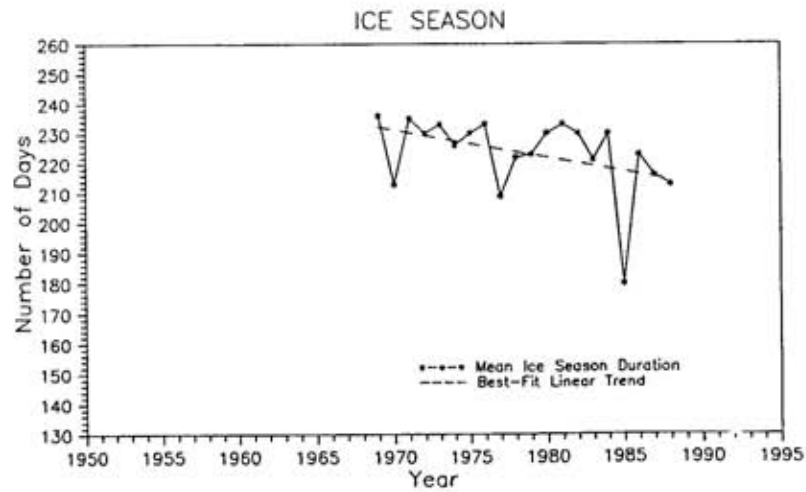
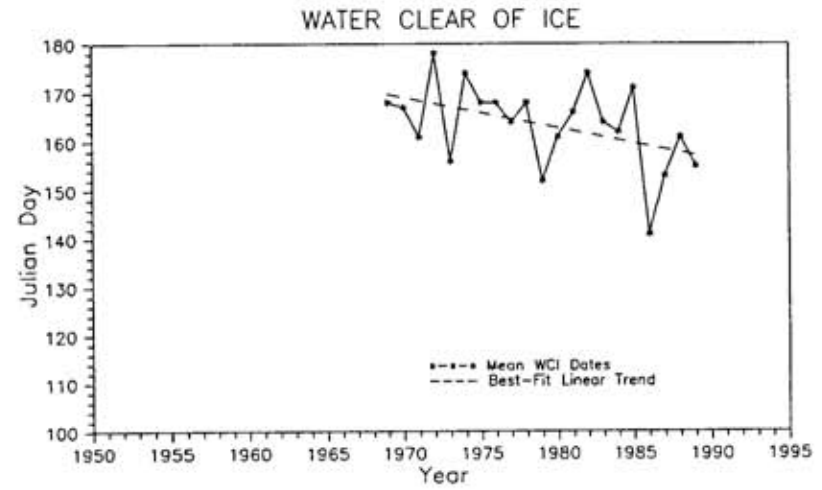
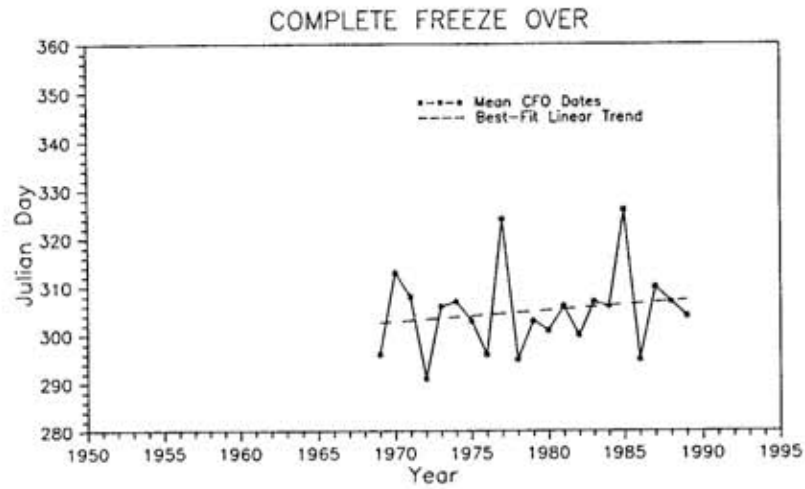


Figure 2h. Lake ice condition time series for the Labrador region.



temperature) and the intercept (lake ice condition), and the standard error of estimate ( $S_e$ ) or the measure of the goodness of fit of the regression line. If  $S_e$  is small, then the predicted lake ice condition dates, or durations, are narrowly scattered about the regression line, indicating that for a given mean air temperature, the predicted lake ice conditions are subject to small sampling error and that 95% of them are within  $2 \times S_e$ . The best equation fits are shown for each lake ice condition. All are significant at the  $\alpha=0.05$  or 95% level and most are significant at the  $\alpha=0.01$  or 99% level better. These tables also show projected regional air temperature increases due to doubling of atmospheric  $\text{CO}_2$ , as predicted by the Canadian Climate Centre General Circulation Model (CCC GCM) (Boer et al., 1990), and the temporal changes that would occur in ice conditions under the calculated statistical relationships.

In most cases, individual regression residuals had normal distributions. Table 7 shows autocorrelation of residuals and the Durbin-Watson statistic for each CFO, WCI, ice season and ice-free season linear regression for all eight lake ice regions. Only one Durbin-Watson ratio suggests moderate autocorrelation in the fitted data. This occurred in the CFO case in the Great Slave Lake region. A Durbin-Watson statistic less than 2.0 corresponds to zero first order autocorrelation while a statistic between 2.0 and 2.5 is inconclusive (Ostrom, 1978). The observed moderate autocorrelation could be random, as it only appears once in any lake ice category. Therefore, the linear regression procedure assumptions are generally satisfied in the observed data.

## Results

A description of the lake ice condition composite for the Great Slave lake region follows. An examination of Figures 2b to 2h and the corresponding Tables 6b to 6h will yield similar descriptions for the remaining seven lake ice regions.

Four lake ice records on or near Great Slave Lake were averaged to yield a common 33 year record for this lake ice region. Long Lake and Frame Lake are small lakes near Yellowknife, NWT while Back Bay and Charlton Bay are located in Great Slave Lake, the former at Yellowknife and the latter at the extreme eastern end of the lake. Temperature indices were derived from the Yellowknife Airport daily temperature record.

Figure 2a shows little change over time in the CFO dates, but there is a general tendency toward earlier WCI dates. The ice season duration curve, as well as the ice-free season duration curve, reflect the moves toward later WCI dates. The regression equations for the Great Slave Lake region are shown in Table 6a. The mean air temperatures, for all periods, are strongly correlated with the lake ice condition dates. The month of October shows the strongest correlation with the freeze-up dates. The period April 21 to June 10 shows the strongest correlation with the break-up dates. For example, a  $2.4^\circ\text{C}$  rise in mean October air temperature due to doubling of atmospheric  $\text{CO}_2$  as predicted by the CCC GCM for the Great Slave Lake area, would result in a delayed freeze-up date for that region of 5.6 days, using the regression coefficient 2.33. A rise of mean April 21 to June 10 air temperature of  $2.9^\circ\text{C}$  would result in a break-up date 8.0 days earlier in the Great Slave Lake area, using the regression coefficient -2.75. An average mean temperature rise during the Fall-Spring seasons of  $2.7^\circ\text{C}$  would result in a reduction in the duration of the ice season of 14.3 days, using the regression coefficient -5.29. Similarly, an average mean temperature rise during the Fall-Spring seasons of  $2.7^\circ\text{C}$  would result in an extension in the duration of the ice-free season of 14.0 days, using the regression coefficient 5.17.

**Table 7.** Autocorrelation of residuals and (Durbin-Watson statistic) from linear regressions for eight lake ice regions.

<u>Lake Ice Region</u>	<u>N</u>	<u>CFO</u>	<u>WCI</u>	<u>Ice Season</u>	<u>Ice-Free Season</u>
Great Slave Lake	33	-0.45 <sup>1</sup> (2.87)	-0.12 (2.24)	0.21 (1.56)	0.05 (1.90)
Southern Yukon	33	-0.20 (2.39)	-0.18 (2.32)	0.26 (1.36)	0.36 (1.27)
Alberta	28	-0.10 (1.99)	0.20 (1.53)	-0.20 (2.27)	-0.02 (1.90)
Saskatchewan - Manitoba	19	-0.08 (1.98)	0.28 (1.43)	-0.30 (2.48)	-0.12 (2.10)
Manitoba-Ontario	33	0.29 (1.42)	-0.11 (2.04)	0.18 (1.63)	-0.14 (2.02)
Northwest Ontario	22	-0.08 (2.12)	-0.06 (1.91)	-0.15 (2.09)	-0.05 (1.92)
Southern Quebec	18	-0.25 (2.21)	-0.21 (2.40)	-0.09 (2.17)	0.06 (1.85)
Labrador	23	-0.14 (2.15)	0.06 (1.87)	-0.39 (1.76)	0.14 (1.62)

<sup>1</sup> significant  $\alpha=0.05$  at 95% level

## Discussion

Eight lake ice regions have been defined and strong statistical relationships have been established between composite lake ice conditions and air temperatures from a station, or stations, in that region. Other internal and external lake factors are also important, as previously described, in the freeze-up and break-up processes. The compositing technique appears, in most cases, to dampen the effects of these other factors and accentuate the regional air temperature signal (Palecki and Barry, 1986).

Regional lake ice conditions appear to be useful indicators of temperature changes in that region during the spring and fall transition seasons. Statistical relationships between ice season and ice-free season and air temperatures averaged over summer, winter or annual periods were weak and in some cases quite poor in most regions. However, strong statistical relationships developed when the air temperatures from the best-fit periods for freeze-up and break-up were combined.

The linear regression results can be used as statistical models to describe lake ice data in terms of mean air temperature, and vice versa. Figure 3a shows the actual and modelled freeze-up (CFO) dates in the Manitoba-Ontario region and the actual and modelled mean November 1 to 20 air temperatures at Winnipeg, Manitoba for the 10-year period 1979 to 1988. The expected CFO dates and air temperatures were derived from the best fit November 1 to 20 linear regression equation for the independent period 1956 to 1978. Similarly, Figure 3b shows the actual and modelled break-up (WCI) dates in the Manitoba-Ontario region and the actual and modelled mean April 1 to May 10 air temperatures at Winnipeg, Manitoba for the period 1980 to 1989. The modelled WCI dates and air temperatures were derived from the best fit April 1 to May 10 linear regression equation for the independent period 1957 to 1979. The close association between air temperature and ice conditions is apparent.

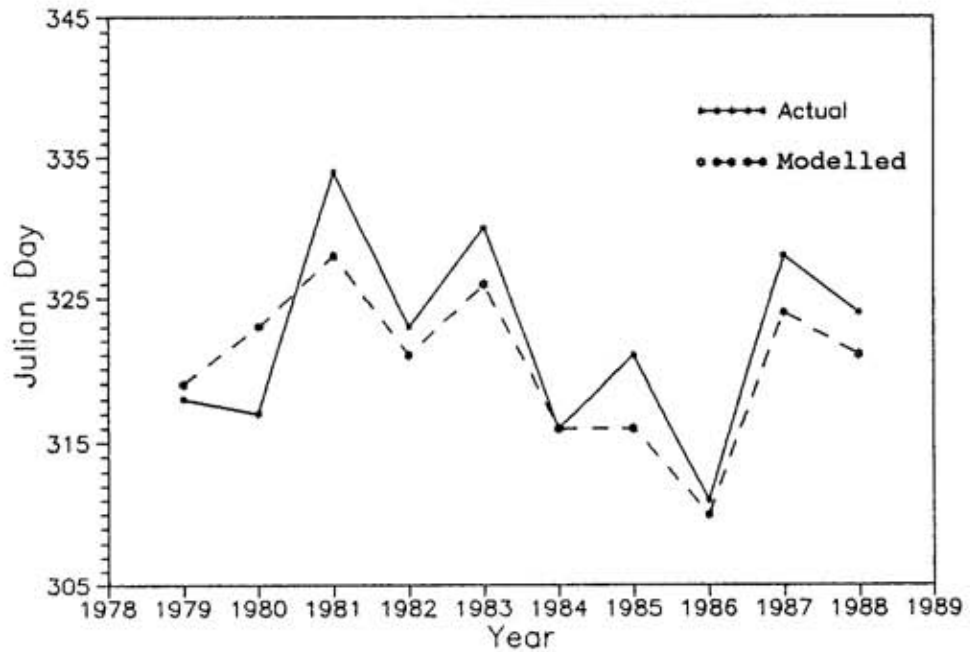
Figure 3c shows the actual and modelled ice season and ice-free season durations in the Manitoba-Ontario region for the period 1979 to 1988. The modelled durations were derived from the best fit Fall-Spring and Spring-Fall linear regression equations, respectively, for the independent period 1956-57 to 1978-79, for ice season, and 1957 to 1978 for ice-free season. The trends towards shorter actual ice seasons and longer actual ice-free seasons during the 1980's is also apparent.

Figure 4a shows the Fall season (September to November) temperature trends, expressed as departures from the 1951-80 average, for three Canadian regions, Northwestern, Western, and Eastern Canada. These regional composites were derived from individual station records from the Historical Canadian Climate Database (Gullett and Skinner, 1992). Each composite is accompanied by a best-fit linear trend line. Figure 4b shows the same regional graphs for the Spring (March to May) season. An examination of the Fall graphs for all three Canadian regions indicates no specific trends over the past 35 to 40 years. The CFO time series of Figures 2a to 2h also show no specific trends towards either earlier, or later, freeze-up for any of the defined lake ice regions. However, an examination of the spring graphs shows distinct trends towards warmer than normal temperatures in both northwestern and western Canada for the entire record, and most notably, since the early 1970's. This is also evident in the spring series, but to a lesser degree, for eastern Canada. The WCI time series of Figures 2a to 2h also indicate equally distinctive trends towards earlier lake ice break-up during the same time period. The trends towards shorter ice seasons and longer ice-free seasons in most regions of the country reflect the earlier spring break-up in response to the warmer temperatures during the spring season, and possibly warmer conditions during the preceding winter season.

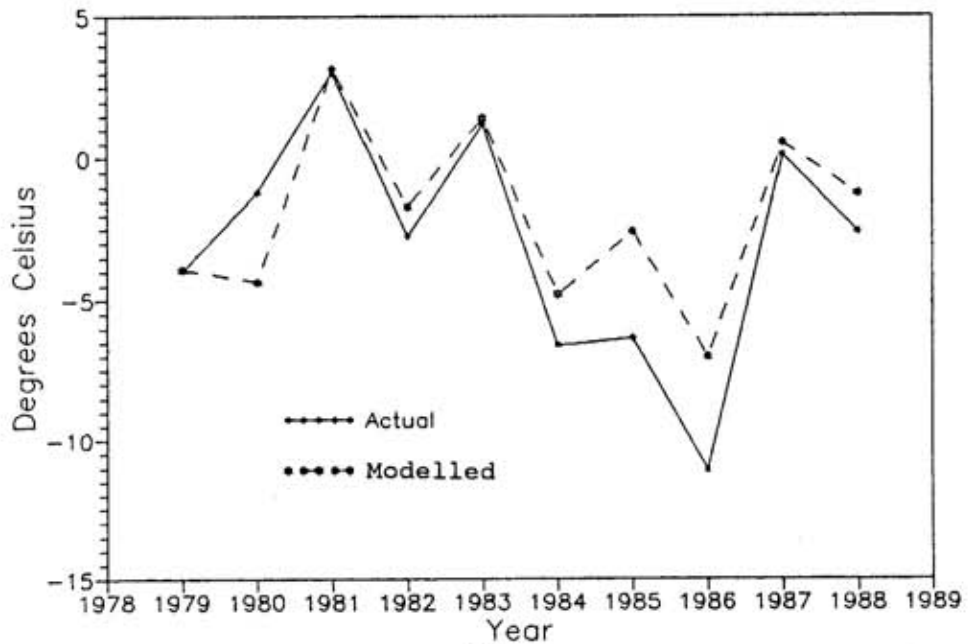


**Figure 3a.** Actual and modelled CFO dates and mean November 1 to November 20 air temperatures for the period 1979 to 1988 as estimated from the best fit regression equation from the period 1956 to 1978.

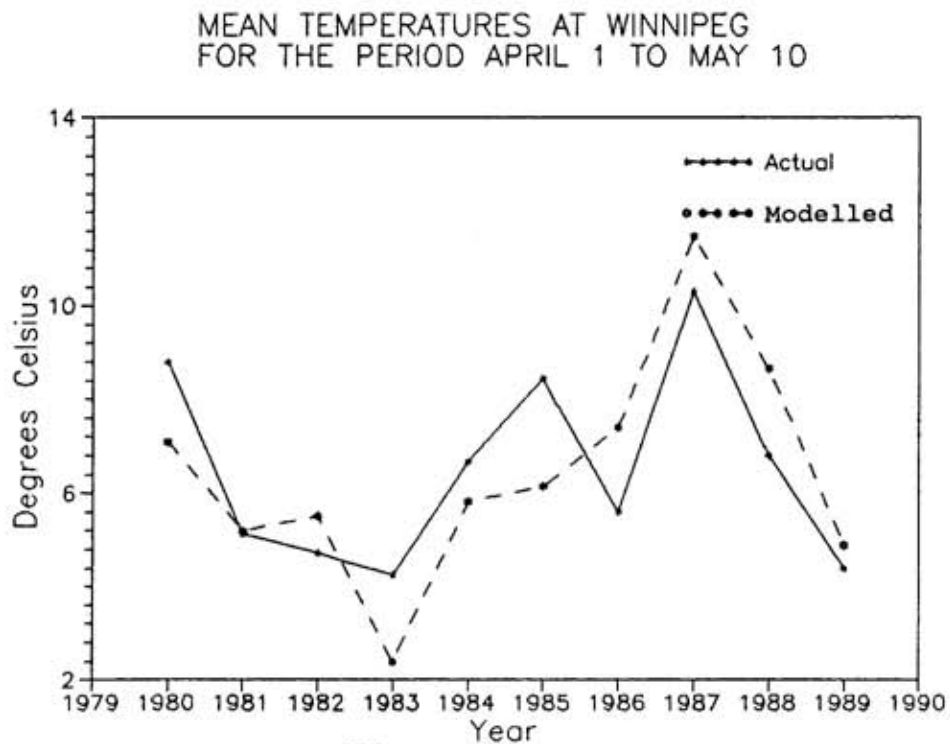
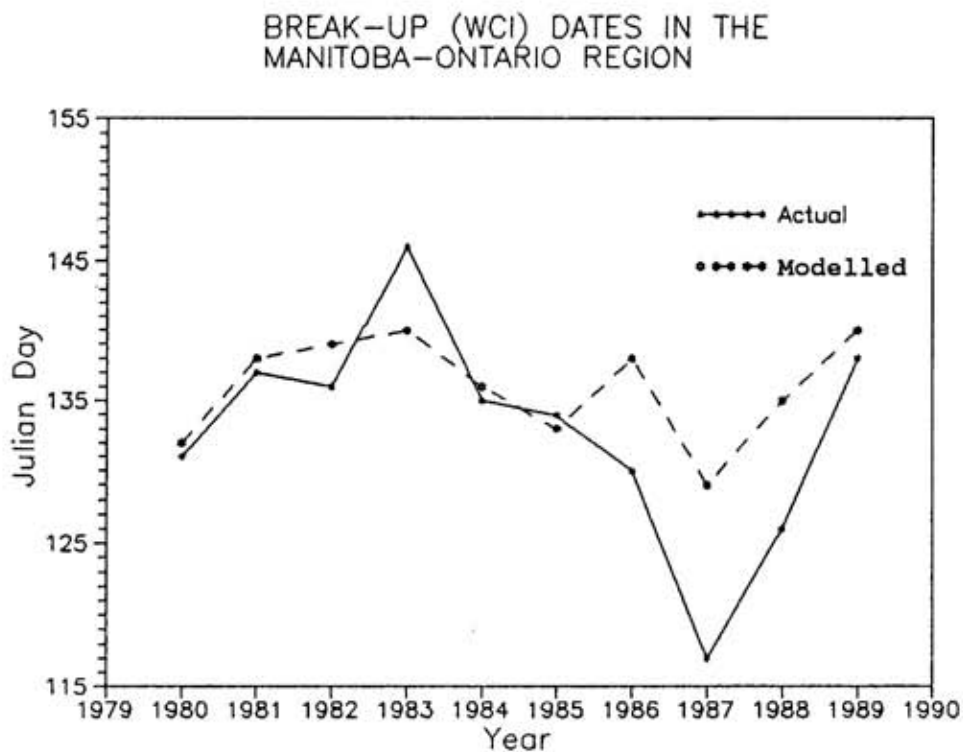
FREEZE-UP (CFO) DATES IN THE  
MANITOBA-ONTARIO REGION



MEAN TEMPERATURES AT WINNIPEG FOR  
THE PERIOD NOVEMBER 1 TO NOVEMBER 20

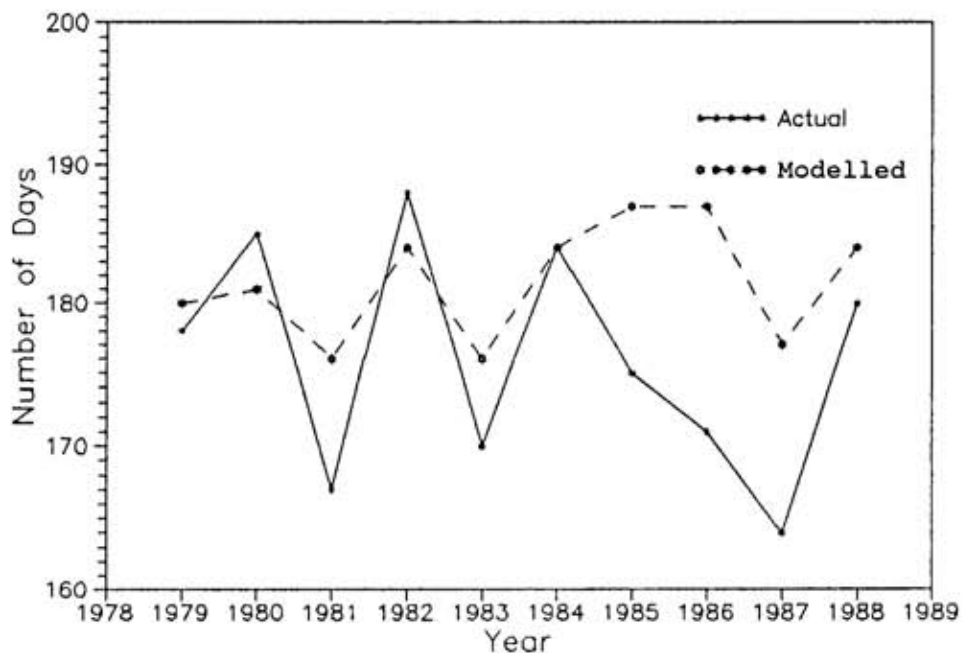


**Figure 3b.** Actual and modelled WCI dates and mean April 1 to May 10 air temperatures for the period 1980 to 1989 as estimated from the best fit regression equation from the period 1957 to 1979.

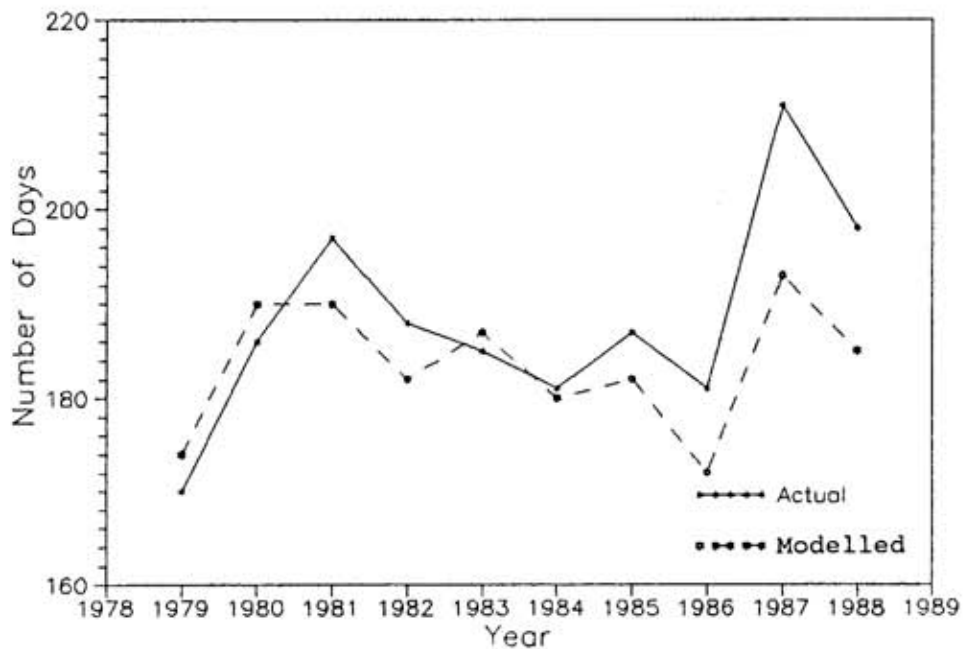


**Figure 3c.** Actual and modelled ice season and ice-free season durations for the period 1979 to 1988 as estimated from the best fit regression equation for the Fall-Spring and Spring-Fall periods, respectively.

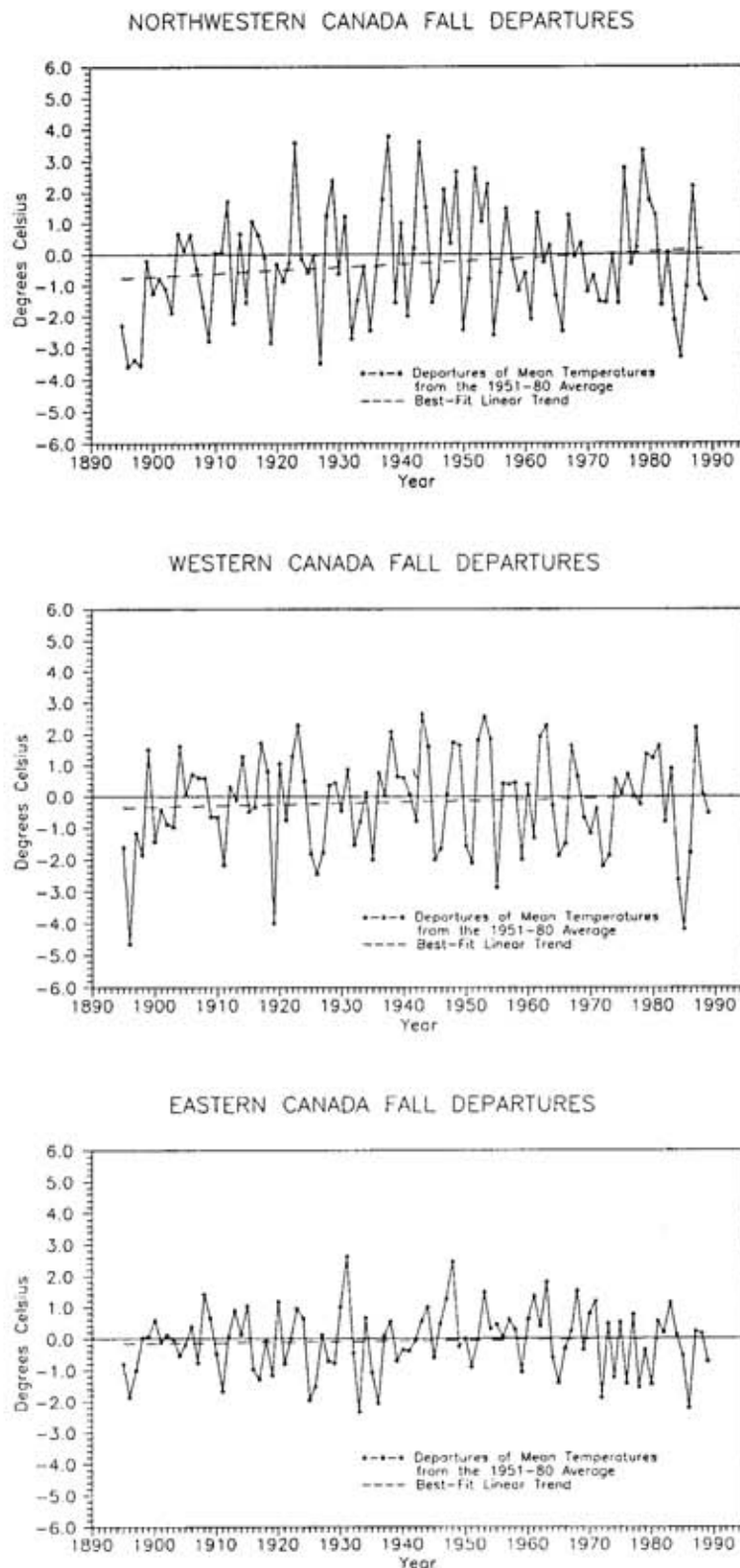
DURATION OF ICE SEASON IN THE MANITOBA-ONTARIO REGION



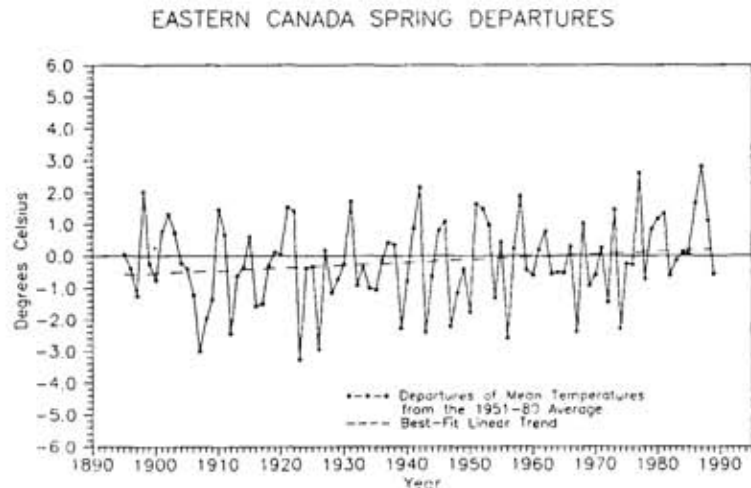
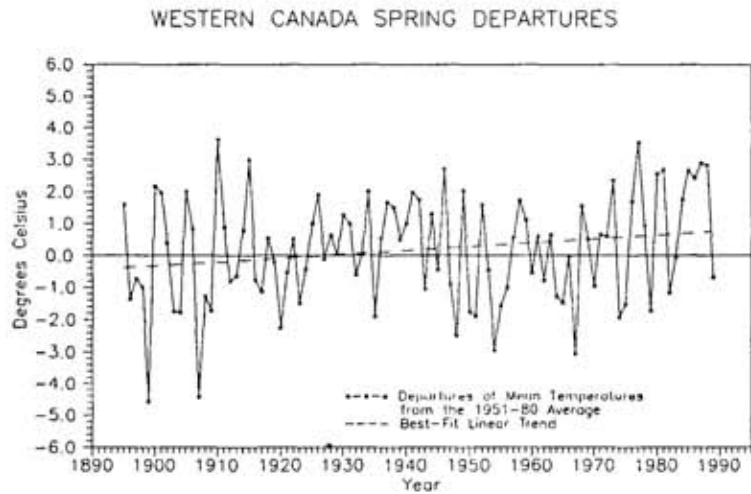
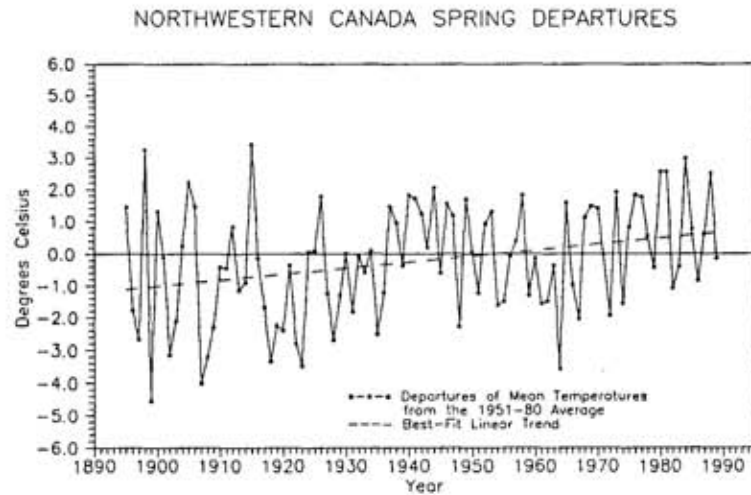
DURATION OF ICE-FREE SEASON IN THE MANITOBA-ONTARIO REGION



**Figure 4a.** Fall season (September-October-November) temperature trends expressed as departures from the 1951-80 average for three Canadian regions.



**Figure 4b.** Spring season (March-April-May) temperature trends expressed as departures from the 1951-80 average for three Canadian regions.





Figures 5a and 5b show changes in fall and spring temperatures, respectively, from the 1959-1973 period to the 1974-1988 period (Kertland, 1990). There is a very small magnitude band, about 0.5°C to 1.0°C, of positive temperature change extending through the Yukon, the Great Slave Lake area, British Columbia and Alberta during the fall season. There is a much broader and higher magnitude band of positive temperature change, about 0.5°C to 2.0°C, during the spring season. It extends from the Yukon, western Northwest Territories, the Prairies, and into Ontario and southeastern Quebec. It peaks over the Prairies and northwestern Ontario. This spring season temperature change would be barely noticeable above the natural variability of break-up dates given the regression coefficients outlined for the six regions in Tables 6a to 6g. However, when combined with the moderate fall season temperature increases, changes in the ice, and ice-free, season durations would be noticeable above the natural variability of the durations.

## Conclusions

The Atmospheric Environment Service (AES) database of observed surface lake ice condition dates has been analyzed to determine the statistical properties of the lake ice data, and to examine the degree of statistical relationship between composite lake ice condition dates and mean air temperatures. Also, composite lake ice condition time series were examined in order to detect any evidence of regional climatic change or variability, and to provide baseline climate/cryosphere relationships upon which climate change impact related hypotheses might be developed.

Eight lake ice regions have been defined and strong statistical relationships have been established between composite lake ice conditions and mean air temperatures from a station, or stations, in that region. Other factors, both morphological and climatic, also determine a lake's response to climate. The compositing technique subdues the effects of other factors and accentuates the temperature signal. Further analysis would incorporate the multivariate effects of changes in these other factors, especially precipitation, snow cover and individual lake heat budget and morphological characteristics, in order to further accentuate the temperature signal.

Regional lake ice conditions appear to be useful indicators of temperature changes in that region during the spring and fall transition seasons. From the developed regression equations, a five to six day change in freeze-up dates for water bodies in the Great Slave Lake represents an approximate 2.5°C change in mean October air temperature. An approximate eight day change in break-up date represents an approximate 3°C change in mean May air temperature. Approximate 3°C increases in average spring-fall and fall-spring mean air temperatures represent approximate 14 day reductions in the durations of the regional lake ice season and ice-free season. A 1°C change in any of the defined lake ice regions represents an average change of about 3 days in both freeze-up and break-up and a change of about 5 days in the duration of the ice and ice-free seasons.

The observed trends during the past decade toward earlier and warmer spring seasons in western and central Canada are reflected in the earlier break-up of the lakes in those regions. No specific trends were seen in either air temperatures or regional freeze-up dates during the fall season in any part of the country. The trends towards shorter ice seasons and longer ice-free seasons in most regions of the country reflect the earlier spring break-up in response to the warmer temperatures during the spring season.

Figure 5a. Changes in fall temperatures from 1959-73 to 1974-88.

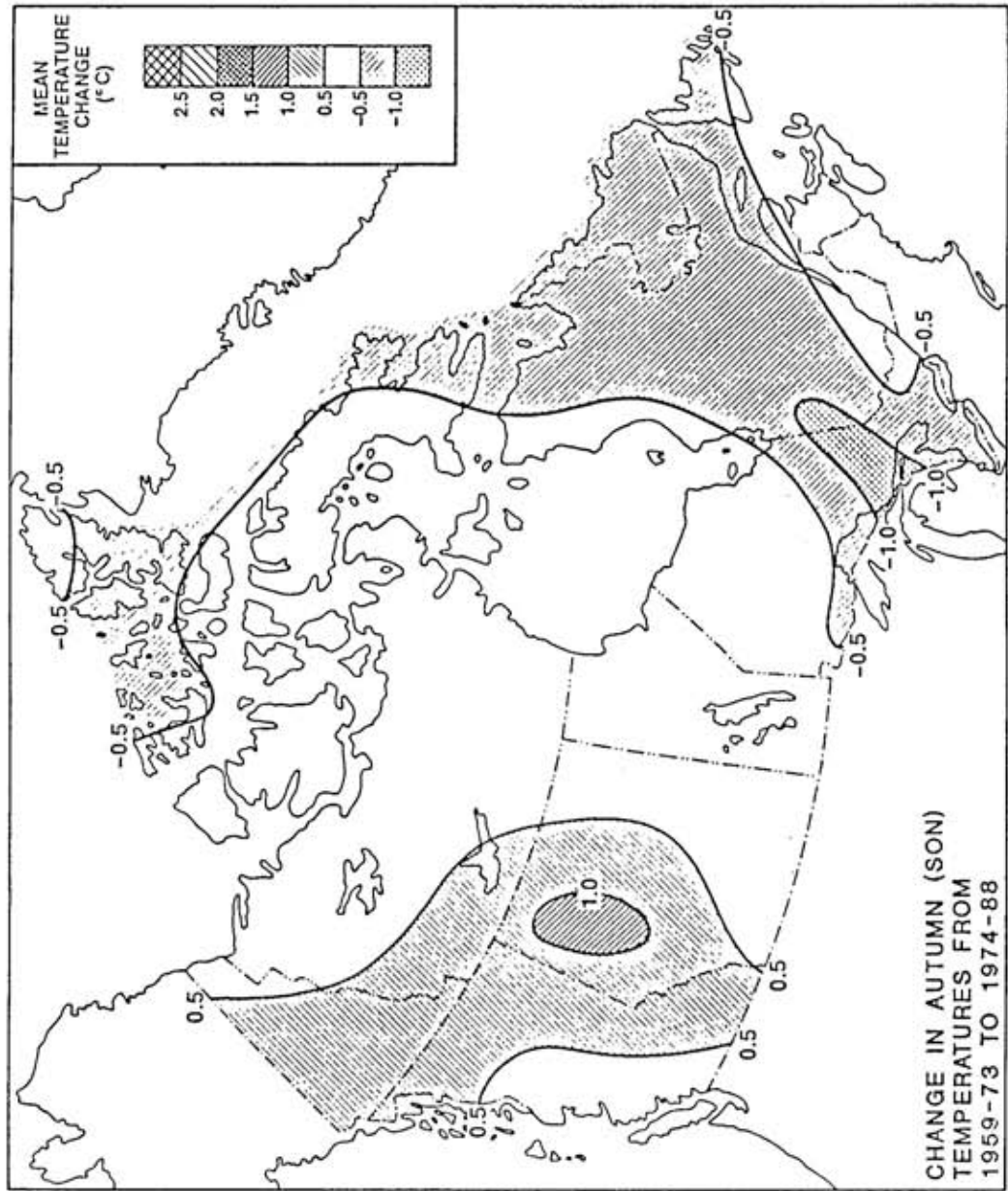
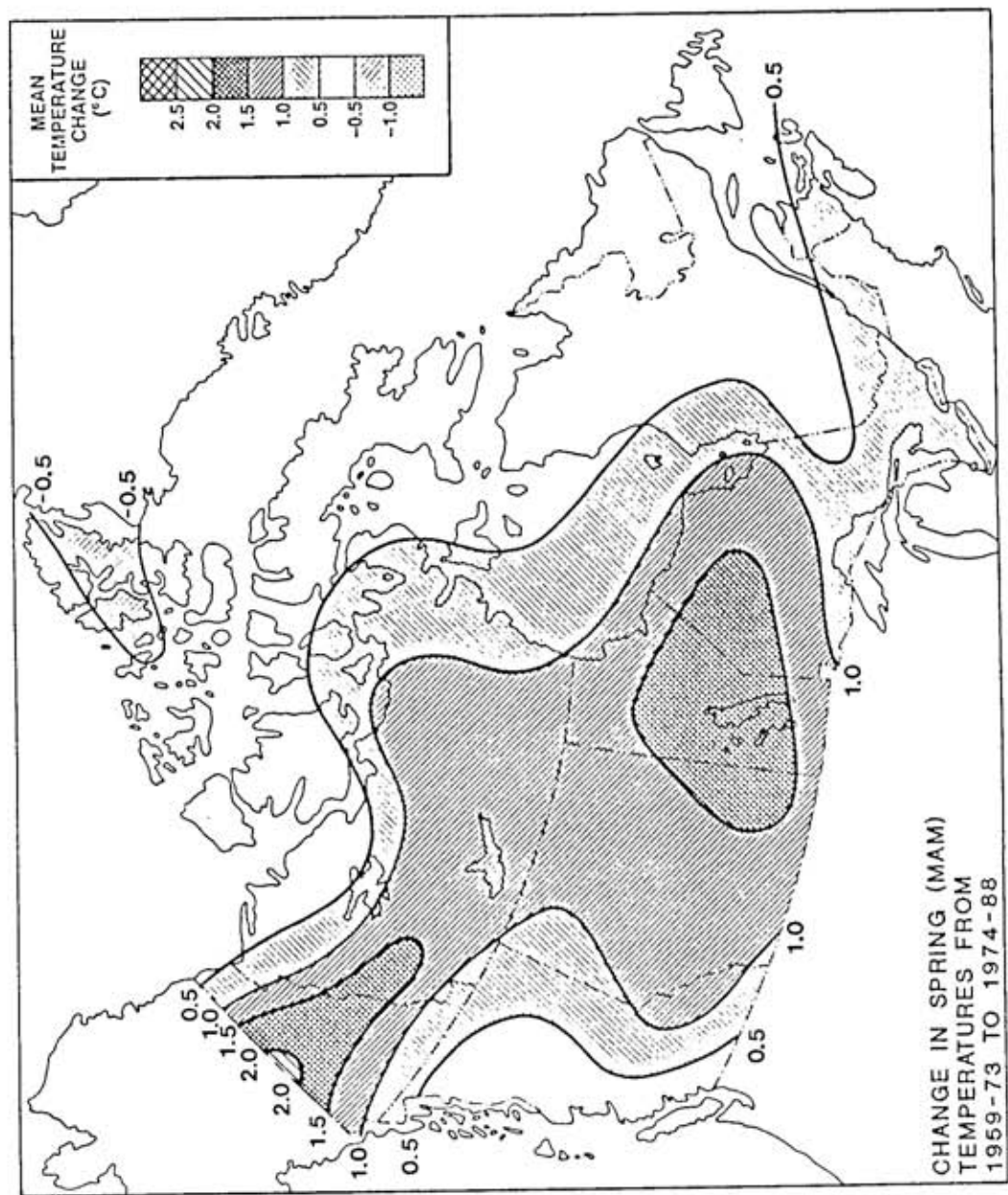


Figure 5b. Changes in spring temperatures from 1959-73 to 1974-88.



These observed surface ice condition data, as in the case of the Great Slave Lake region, can be used as ground truth for the more spatially and temporally consistent satellite monitoring of lake ice freeze-up and break-up and lake surface temperatures. Also, they can be used to assist in model development for use with the analysis of other lakes where there are either no surface data, such as Great Bear Lake, or unreliable data, such as Lake Athabasca. Subsequently, an evaluation can be made of the impact of changing regional environmental conditions on lake ice conditions in that region.

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# Data Needs for Detection of Climate Change: Some Challenges

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## Introduction

There are many uncertainties in the prediction of future climate changes, particularly with respect to timing, magnitude, and regional patterns of change. Monitoring of the climate system through systematic observation and analysis of climate-related variables on a global, national and regional basis is one contribution to improving our predictive capability. The ability to identify a significant change in a climate variable is an integral part of this monitoring.

The highest priority science question identified by the EOS Panel on Physical Climate and Hydrology (1991) was to determine how the global and regional storages and fluxes of water, including snow cover in northern latitudes, will interact with a change in climate. The IPCC (Houghton et al., 1990) noted that snow, ice and glacial extent are key variables in the global climate system and that accurate information on cryospheric changes is essential for full understanding of this system. Cryospheric components are often very sensitive indicators and integrators of basic climate elements, such as precipitation, temperature and solar radiation. Information on the cryospheric system must come from conventional and remotely sensed data. Key elements must be defined which should be routinely monitored on a regional and global basis. But what are the challenges facing us for these to be used effectively for monitoring and detection? This paper will focus on snow cover as an example of a cryospheric element for monitoring climate variability and change. A challenge to the statistical community on how we might properly assess variability or detect change in this element naturally evolves.

## Snow Cover an Indicator of Climate Variability and Change

The development of a consistent, compatible data series for spatial and temporal analyses of snow cover variability is critical. Extraction of a data set complete enough to address the problem at hand can be time consuming and costly; this is even more critical when using satellite data since the volume of data to be processed and analyzed is much greater than conventional data.

Inherently we feel that cryospheric changes should reflect changes in climate, but in the case of snow cover what elements should we monitor? Should it be the:

- change in snow water equivalent at peak accumulation for selected locations?
- change in date of the peak accumulation?
- northward retreat of the snowline on a given date?
- change in the spatial distribution of snow cover over a region?
- change in number of days of continuous snow cover?
- change in date of the beginning or end of continuous snow cover?
- change in the date of the onset of spring melt?
- change in the frequency of the number of melt events during the winter season?
- mapping and identification of temporal outliers from the main accumulation period?



These are only some of the potential "signals" which one might wish to monitor. However, each requires data to be analyzed differently. Current cryospheric data sets are a mix of ground measurements and information derived from satellite remote sensing. For change analysis the frequency of collection, archiving and processing of conventional or satellite data into product information may be daily, weekly, monthly or be limited to a set date. The question of integrating ground data collected at specific locations and on certain dates with remote sensing information must be addressed by researchers as they develop a strategy for using snow cover as an indicator of climate variability and change.

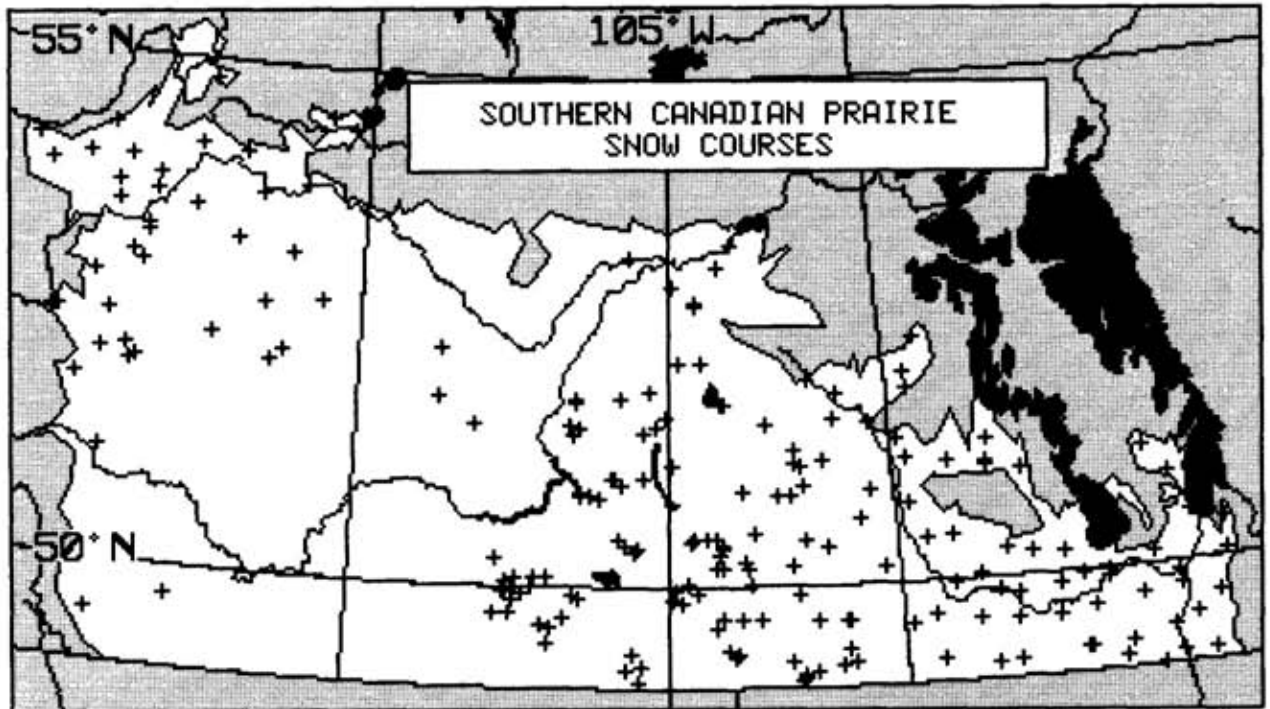
### **Conventional Snow Cover Data**

Ground data may be collected at specific sites on set dates every year or be collected at an irregular network of stations on varying dates; stations unfortunately are under constant pressure of change as they are opened and closed to meet operational needs. Yet a time series of consistent and compatible data must be prepared for assessment of variability and change. In this paper, particular focus will be on the Canadian prairies and the ground and satellite information which one might use to study variability and change.

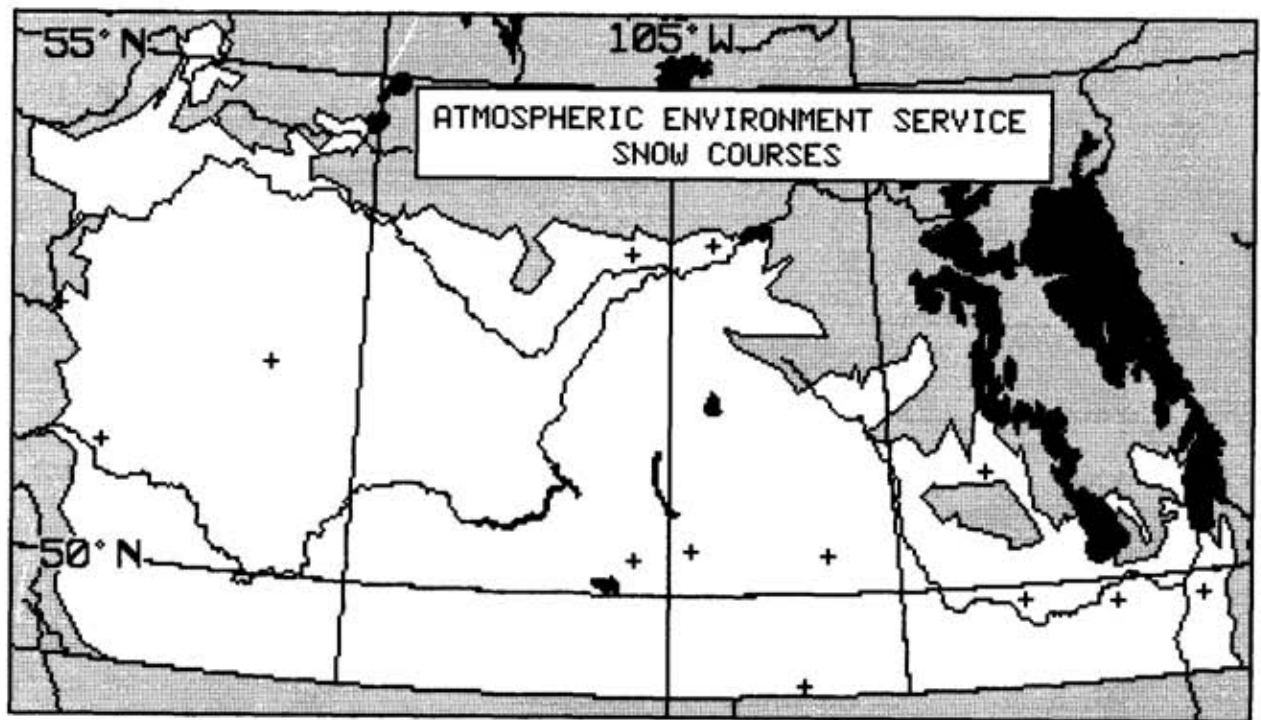
The development of a snow survey network and the frequency and accuracy of associated measurements must be related to its purpose. Most snow courses are established to support water resources programs. They are established to obtain an index of snow water equivalent over an area. For this purpose, the equipment, procedures and siting of the snow course should remain consistent over time. Unfortunately, some or all of these factors may change over the years; yet the data continue to be used in the same manner, uncorrected, with users ultimately questioning their validity. Absolute estimates of areal water equivalent can only be obtained after allowance is made for instrumental, observational and siting biases and after the sampling network has been designed to reflect areal snowcover variability. These issues are discussed in Goodison et al. (1987). One must be extremely careful in using the snow course data for time series analyses and for developing or validating remote sensing algorithms. It will be a challenge to determine how these data can be used for statistical analyses of variability and change.

The snow survey network in southern Alberta, Saskatchewan and Manitoba is shown in Figure 1. It is evident that there is high spatial variability in the areal coverage, with large areas having no snow courses at all. The analytical problem is compounded when one realizes that the courses operated by Alberta Environment are measured at the end of February and March whereas those operated by Saskatchewan and Manitoba provincial agencies are measured during the second week of February and March. Temporally the data are incompatible. The only snow course data available on a weekly or bi-weekly basis are from those stations operated by the Atmospheric Environment Service; those few stations are shown in Figure 2. Thus an improved temporal record, and one suitable for time series analysis, results in poor spatial coverage. In all cases in this region, annual changes in land use will adversely affect any meaningful time series analysis of variability and change. In all cases one assumes that the snow course network will be sampled at each scheduled survey. This is extremely important for analyses of variability and change; however, in 1988, a low snow year, the surveys were not conducted in Saskatchewan and Manitoba. The low snow values were not recorded; "missing", not the assumed zero value, is the official observation. This type of problem is critical when trying to establish current variability and ultimately detect statistically significant change; the low values in a time series must be known. Despite these problems, however, the snow course network provides the only conventional, or ground data, that can be used to study the distribution and change in snow water equivalent over the region. Snow course records also provide

the only information on snowpack structure, which are useful in the development and interpretation of remote sensing snow cover products.

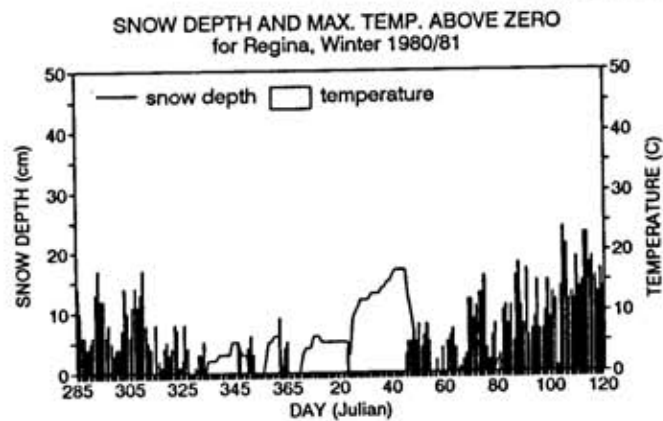


**Figure 1:** Snow courses operated by federal and provincial agencies in southern Alberta, Saskatchewan and Manitoba.

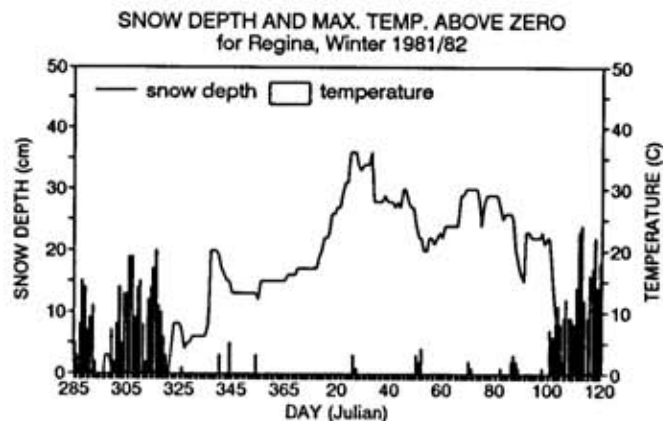


**Figure 2:** Snow courses operated by the Atmospheric Environment Service in the southern Prairie region.

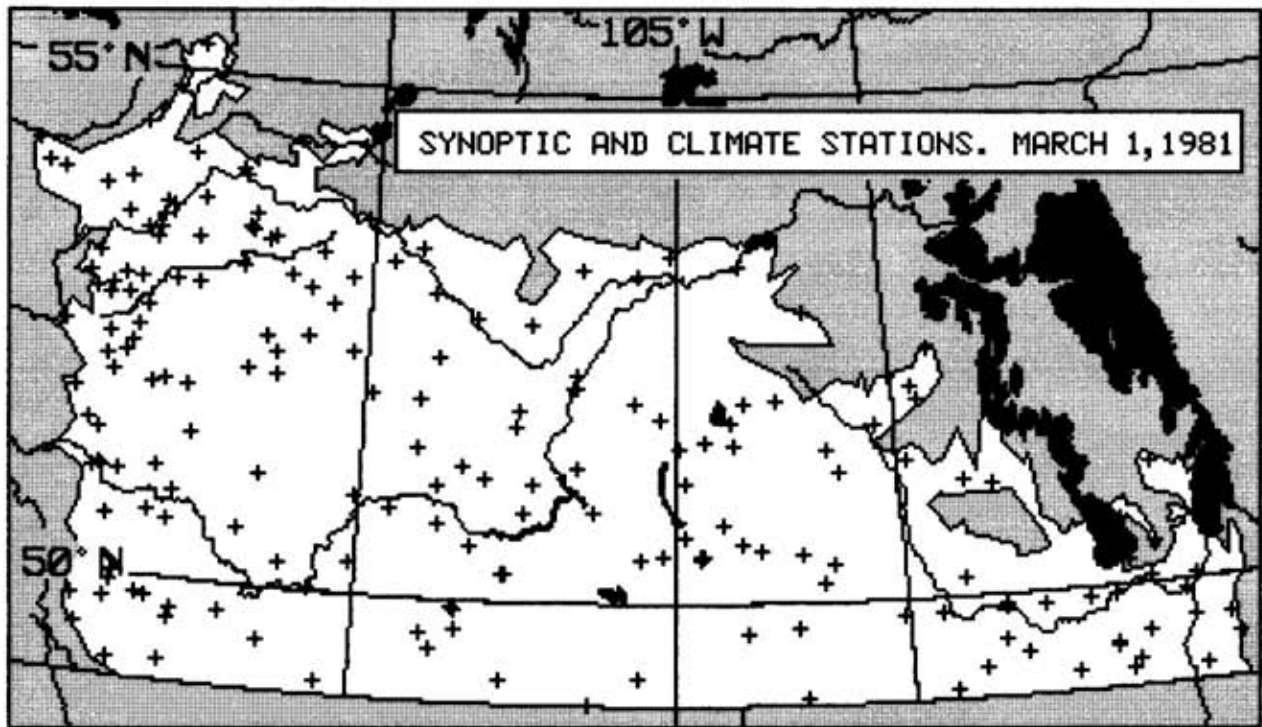
An alternative source of snow cover information is the snow depth observation made daily at all climate (since 1980) and synoptic stations in the region. Data from these stations still form the basis of much of our analysis of snow cover variability and change on a regional basis. Figures 3 and 4 compare snow cover for Regina, Saskatchewan for the 1980-81 and 1981-82 winter seasons as depicted by the daily snow depth recorded at the synoptic station. Occurrences of daily maximum temperatures above 0°C are also plotted and provide an indication of possible melt events. The two winter periods are quite different, with 1981-82 exhibiting a more continuous snow cover with few winter melt events. A time series of these archived data can contribute directly to addressing several of the questions noted above. One can compare differences in the amount and date of maximum snow accumulation, the number of days of continuous snow cover, differences in the onset and ending of continuous snow cover, and so on. However, there is no information about spatial variability of the snow cover. Other synoptic or climate station data would be required in order to derive spatial snow cover information for a region of interest; the representativeness of the information would depend upon the density of the station network. Figure 5 depicts the distribution of synoptic and climate stations over the Canadian Prairie region where snow depth measurements were recorded on March 1, 1981. Although there are 190 stations within the study area, their spatial distribution is by no means even over the region; reliable information about the spatial variability of snow depth would be limited in the areas where the stations were widely spaced.



**Figure 3:** Daily snow depth (cm) and maximum air temperature (above 0°C) for Regina Saskatchewan, October 1980 to April 1981.



**Figure 4:** Daily snow depth (cm) and maximum air temperature (above 0°C) for Regina Saskatchewan, October 1981 to April 1982.



**Figure 5:** Distribution of Atmospheric Environment Service (AES) synoptic and climate stations with recorded snow depth measurements on March 1, 1981.

### **Contribution of Remote Sensing Information**

On a global basis, conventional data are difficult to use, if only because they are not readily available. Remote sensing does offer an alternative source of information. To get a better global scale perspective of snow extent one would more likely use the NOAA/NESDIS weekly Northern Hemisphere snow cover maps (snow/no-snow) derived from visible satellite imagery (e.g. NOAA, GOES, METEOSAT) (Wiesnet et al., 1987). Recent re-analysis by Robinson et al. (1991) of this product to assess the change in Northern Hemisphere snow cover showed a decrease in snow cover of about 8% from the mid-1970's to the late 1980's, with snow extent being especially low in spring since 1987. However, these products can only address the issue of snow covered area, and its variability, not the water equivalent or depth of a snow cover.

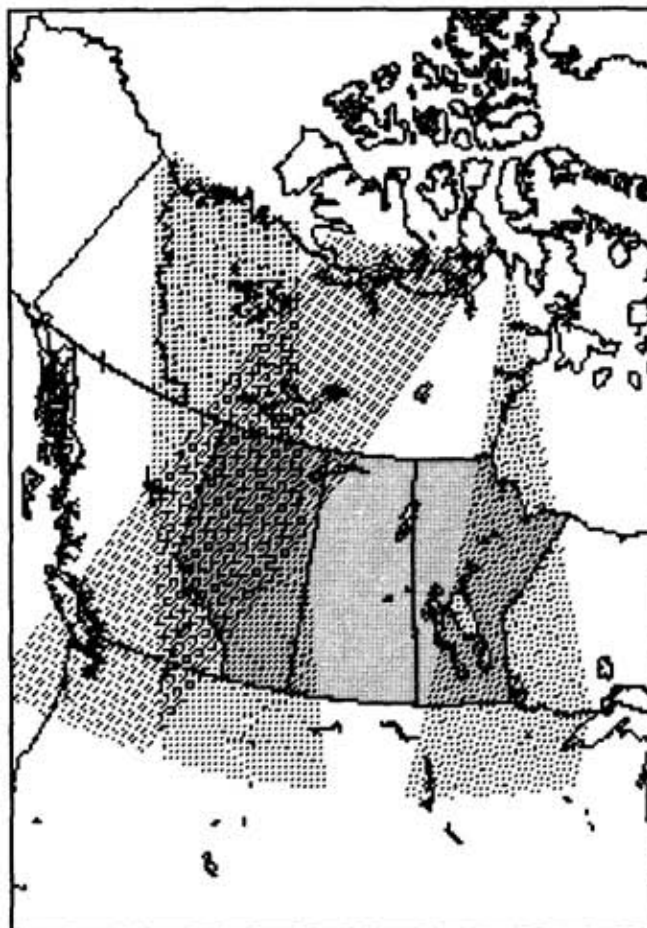
One of the limitations of the NOAA weekly chart is that it is a weekly composite map derived from visible satellite data; cloud cover will preclude accurate assessment of the snow extent on any given day and sometimes it is necessary to use the extent derived during the previous week. This is a natural limitation of all optical data for operational snow cover mapping. A major advance in the remote sensing of snow cover has been the development of algorithms to determine snow depth, water equivalent, areal extent and snow state (wet/dry) using microwave satellite sensors having all-weather capabilities (Hallikainen, 1992; Goodison et al., 1989; Goodison, 1989; Chang et al., 1989; Goodison et al., 1986; Kunzi et al., 1982).

Algorithms to derive snow water equivalent (SWE), areal extent and snowpack state (wet/dry) using satellite passive microwave data were developed from intensive airborne studies over the Canadian prairies and tested using NIMBUS-7 SMMR and SSM/I data for this mid-continent region (Walker and Goodison, 1992; Goodison et al., 1990; Goodison, 1989; Goodison et al., 1986). Although this research was initiated to support



hydrology, the derived snow cover information is now being assessed for its potential in the study of climate and global change. The use of snow cover as an indicator of climate change and the effective use of passive microwave data for such analyses are being assessed in the recent effort of the Canadian Climate Centre to produce a snow cover time series for the Canadian Prairies for the SMMR period (1978-1986) as a contribution to the International Space Year (ISY).

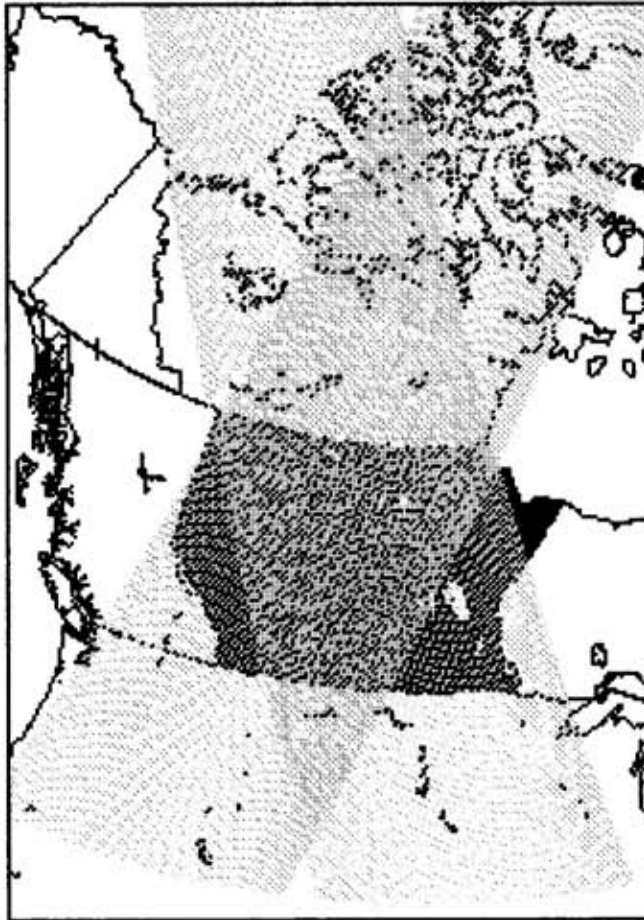
Although a reliable, accessible archive of passive microwave data is readily available since the launch of NIMBUS-7 SMMR in 1978, there are some practical limitations to the production of snow cover time series, even for relatively small regions. The Canadian open prairie region spans a longitude of only  $18^{\circ}$  and covers an area of only  $512,560 \text{ km}^2$ . However, the SMMR swath was only  $800 \text{ km}$  wide and due to satellite power constraints the sensor was operated every second day. Data from two or three successive operational periods were required to provide complete coverage of the study area, resulting in a composite map of snow cover for a three to five day period (Figure 6). One is then forced to analyze a weekly composite of snow cover rather than a daily product. Changing weather patterns over the five day period could produce changes in the SWE and distribution of snow cover and changes in the occurrence and distribution of melt areas or wet snow (resulting in an underestimate of SWE and snow covered area from the microwave data) which would produce a non-representative picture of snow cover across the region for the period.



**Figure 6:** Nimbus-7 SMMR swath coverage for a single day (DAY 1) over western Canada. For complete coverage over the region, data were required from orbits over a 5-day interval (DAYS 1,3,5).

The situation is much improved for the operational DMSP SSM/I data, which has been available since 1987, as the  $1400 \text{ km}$  swath provides daily coverage of this particular study area, eliminating the problem of compositing and interpretation of potential artificial variations in snow cover (Figure 7). Daily coverage of the study region is preferred, allowing easy comparison with conventional ground based data from synoptic and climate stations and snow courses. Figure 8 illustrates the distribution of pixel values of snow water equivalent over southern Saskatchewan; also plotted are airborne gamma measurements of SWE averaged over  $20 \text{ km}$  flight line segments. It is evident that the gamma data provide more information on SWE than point measurements, but the satellite data are capable of providing even more spatial information. It must be noted though that

exact pixel locations change from orbit to orbit. For statistical analysis of such time series, one might consider gridding the satellite data to a fixed area on the ground before analysis. However, use of data from SMMR and SSM/I to produce a consistent, compatible time series of snow cover from year to year will be difficult and must be done with care and caution; products from the two satellite sensors cannot be directly compared unless the same compositing procedure is used for both data sets.



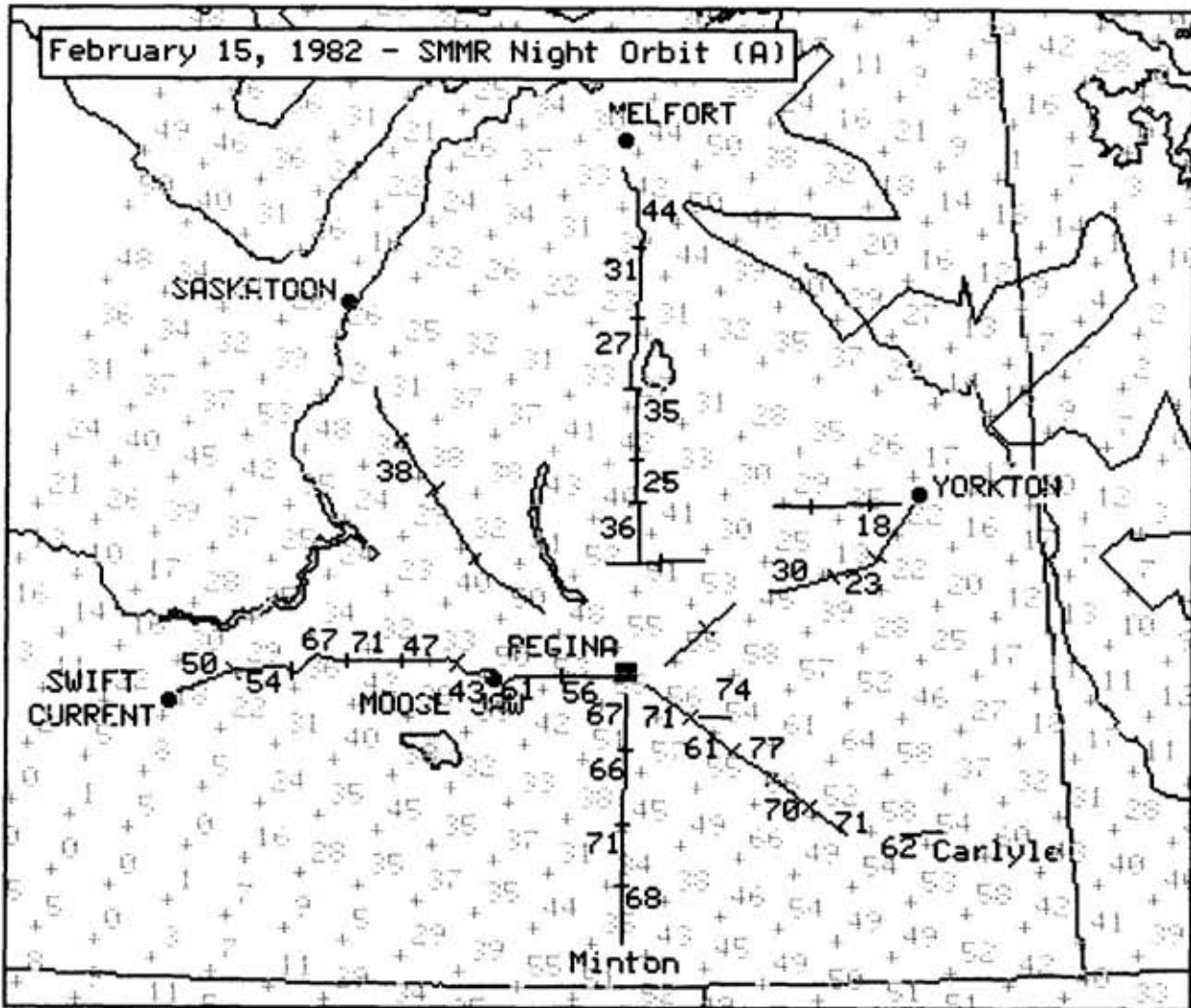
**Figure 7:** DMSP SSM/I swath coverage over western Canada. Complete coverage of the Canadian Prairie region can be obtained on a daily basis using one ascending and one descending pass.

As noted, NOAA AVHRR data are used to produce maps of snow cover extent. However, cloud cover can limit their effectiveness in determining the actual dates of the beginning or end of snow cover. This has been particularly true for the fall period. It was expected that passive microwave data could overcome some of these limitations. Success has been achieved for open areas when there is dry snow (Goodison, 1989; Goodison et al., 1986). Figures 9 and 10 show the regional snow water equivalent as derived from SMMR data for the first week of March in 1981 and 1982. The large difference in areal coverage and distribution between the two years is readily evident. It is this type of information, put into a weekly time series for every snow cover season, that can be assessed for variability and change analyses. It complements the information available for individual stations (see Figures 2 and 3). It provides the spatial distribution of SWE which is not available from conventional or other remote sensing observations. For example, Figure 3 shows there is no snow cover at Regina; Figure 9 illustrates how large an area is actually devoid of snow. Figure 4 provides a detailed daily time profile of snow depth at Regina, suggesting extensive cover during the 1981/82 winter; Figure 10 shows that in fact the area of peak accumulation was centred near Regina and other areas within the region had considerably less snow.

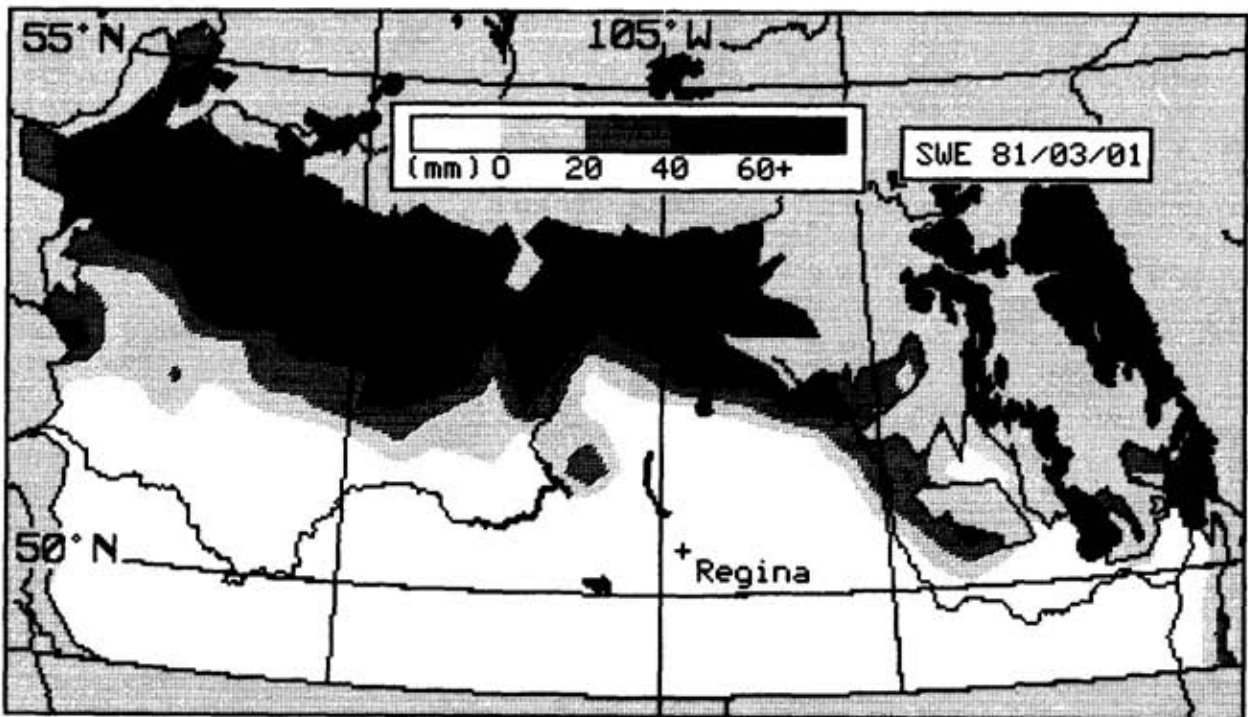
Snow cover maps derived from passive microwave data do offer the great advantage of assessing changes in areal snow cover water equivalent during the winter and from year to year, allowing for the current limitations noted above. Figures 11 and 12 compare the snow covered area for the first weeks of February and March for 1979 to 1986. Early



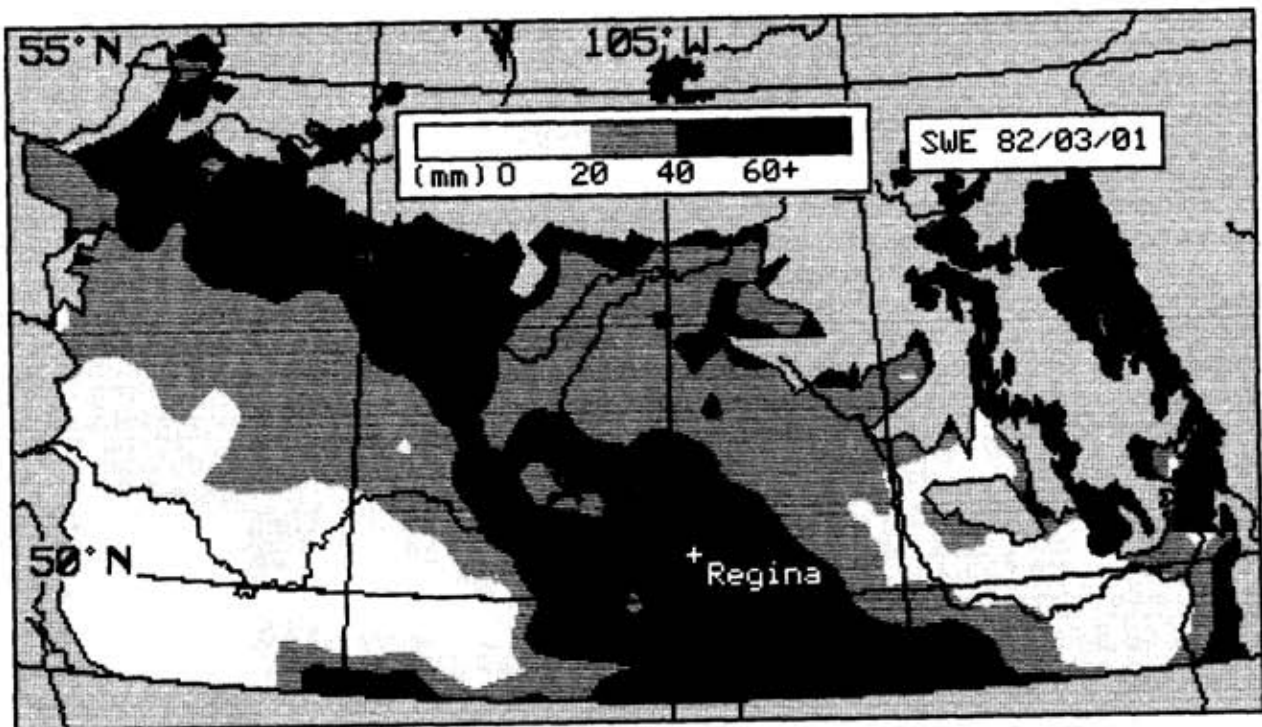
February shows the tendency to have a higher percentage of snow covered area over the entire prairie area than during early March. The March survey, however, may give a better indication of the potential for soil moisture recharge and runoff than the earlier survey alone. For this eight year period, no clear trend emerges with respect to snow cover over the Prairies; the SMMR record is just too short to draw conclusions about climate variability and change. When properly combined with SSM/I and conventional records, the record can be extended to allow for a more accurate analysis of temporal variability.



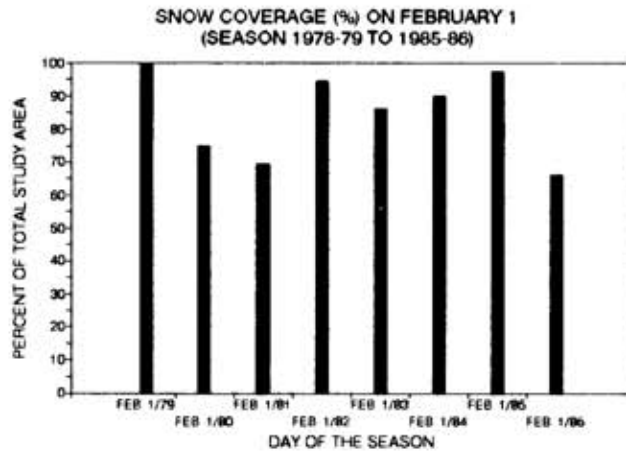
**Figure 8:** Snow water equivalent (mm) derived from SMMR satellite data for southern Saskatchewan on February 15, 1982. Airborne gamma estimates (mm) are plotted in black.



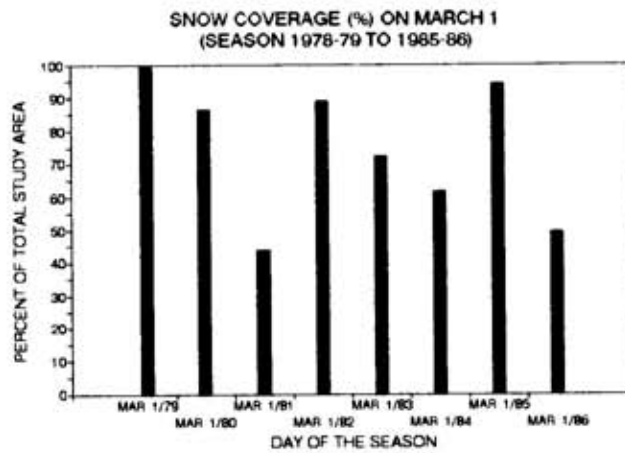
**Figure 9:** Distribution of snow water equivalent (mm) over the Canadian Prairies for March 1, 1981 derived from Nimbus-7 SMMR data.



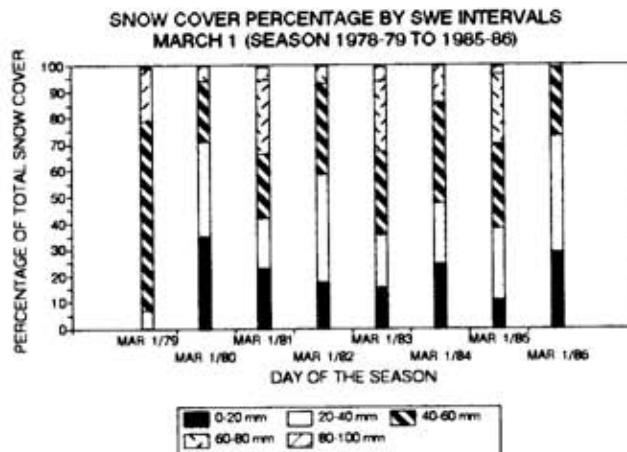
**Figure 10:** Distribution of snow water equivalent (mm) over the Canadian Prairies for March 1, 1982 derived from Nimbus-7 SMMR data.



**Figure 11:** Snow cover as a percentage of total area of the Canadian prairie region for the first week of February, 1979 to 1986.



**Figure 12:** Snow cover as a percentage of total area of the Canadian prairie region for the first week of March, 1979 to 1986.



**Figure 13:** Snow cover as a percentage of total snow covered area for 20mm intervals of snow water equivalent, for the first week of March, 1979 to 1986.

Another substantial contribution of the passive microwave derived snow cover product is its ability to map and compute the areal coverage of SWE, thus allowing comparison of the amount of water available from year to year. Figure 13 compares the area covered by different amounts of snow water equivalent, in this example using 20 mm increments. For example, 1986 was characterized by both low areal extent and low water equivalent; 1981 on the other hand had low areal extent, but had a considerably higher water equivalent where there was snow. Again no discernible trend is evident over the SMMR period for areal coverage of a particular amount of water equivalent. The satellite data are the only viable data source for assembling these statistics on snow cover and amount on a weekly basis. The conventional record cannot provide the spatial variability for such an analysis and only snow courses can provide a measurement of SWE; there are too few of these sites to construct an accurate areal picture on a weekly time scale throughout the season.

## Conclusions

Accurate information on cryospheric changes is essential for a full understanding of the climate system. Key elements must be defined which should be routinely monitored on a regional and global basis. Discussion of the questions raised in this paper would provide a start in defining the elements which could be monitored and in identifying appropriate data sources for obtaining information on the elements. Ideally both conventional and remotely sensed data will be used. A challenge to the statistical community on how we might properly assess variability or detect change in these elements naturally evolves.

Current cryospheric data sets are a mix of ground measurements and information derived from satellite remote sensing. For change analysis, the frequency of collection, archiving and processing of conventional and satellite data into product information may be daily, weekly, monthly or be limited to a set date. Ground data may be collected at specific sites on set dates every year or be collected at an irregular network of stations on varying dates. To complicate matters stations are opened and closed almost at random with generally no concern over their contribution to the cryospheric data they contribute. Yet a time series of consistent and compatible data must be prepared for assessment of climate variability and change. The use of spatially averaged information derived from remote sensing further challenges the analysis of change.

Following are some of the challenges and some suggestions for continued research, if we are to build a cryospheric, and especially a snow cover, data set for analysis of climate variability and change.

1. Snow cover (and other cryospheric) elements which should be monitored to study climate variability and change in Canada must be identified.
2. A consistent, compatible data series for the spatial and temporal analysis of snow cover in Canada must be developed.
3. To achieve the above, all available conventional snow depth observations must be put into the digital archive. The Canadian snow course data must be digitized; the snow course records since 1986 must be published. Stations with a "long-term" homogeneous data record must be identified.
4. A remote sensing time series of snow cover is too short for statistical analysis and detection of climate change. Even the combined existing conventional snow depth and snow course record may be too short.

5. Methods have to be developed to integrate conventional and remotely sensed data to create as accurate a representation of snow cover as possible.
6. Methods must be developed to combine conventional and remotely sensed time series to extend the length of the snow cover record. For example, how do we combine conventional ground observations with the NOAA snow cover product, the SMMR record, and the SSM/I record to create a time series of snow extent or water equivalent over the last fifty years?
7. There must be investigation of the appropriate temporal and spatial scale of data integration for analysis of climate variability and changes.
8. There must be continued development of algorithms to derive snow cover and other cryospheric elements from remotely sensed data, especially passive microwave satellite data.
9. Continued investigation of compositing procedures, especially using remotely sensed information, for the production of daily, weekly or monthly time series for the assessment of variability and change is essential.
10. In concert with statisticians, an assessment of available statistical procedures to analyze snow cover information in the context of climate variability and change detection is required. New techniques for the analysis of "short" spatially averaged environmental data sets will likely be required.

The above thoughts are by no means all inclusive. It is clear, however, that the development of a strategy for using such cryospheric information as an indicator of climate variability and change is a fundamental challenge for climate, cryospheric and statistical researchers.

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## **The following are minutes of the working groups**

We have made tape recordings of these working groups.

We will be asking a key representative of each working group to go over the recordings and transcribe salient points that will add dimension to these minutes. These transcriptions will be published as an addendum to this document in the near future.

# Working Group #1: Remote Sensing

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*R. Leconte, Rapporteur*

Satellite data are well suited, and in some cases the only means, for the monitoring of snow at the regional to continental to global scales. However, many important issues, in particular those pertaining to integrating remotely sensed data with conventional surface data, and the accurate detection of snow parameters at scales suitable for global monitoring, further need to be addressed. The working group focused on issues related data integration, snow detection, and algorithm development and sciences issues.

## Data Integration

A major recommendation from the Snowwatch'85 meeting pertained to the integration of conventional and remotely sensed data to create more complete data sets. Since then, some progress has been made, and in some cases, the merging of these data sets is now being done operationally. For example, the national Operational Hydrologic Remote Sensing Center of the national Weather Service now routinely collect and merge together AVHRR, airborne gamma, and ground data of snow cover to generate maps of snow cover extent and water equivalent for selected basins in the U.S. and in Canada. However, the level of effort needs to be increased, including larger coverage, increased frequency for generating products, and exploring the synergism of combining various types of remotely sensed data.

### *Recommendations*

- 1) To increase the level of effort in integrating remotely sensed and conventional data sets for calibration and validation of snow retrieval algorithms and for addressing key scientific issues, e.g. snow albedo variations with snow depth.
- 2) Considerable research has been done in developing algorithms to extract snow water equivalent and depth from remotely sensed data, e.g. passive microwaves. However, the potential of merging different types of remote sensing data has not been significantly exploited. It is recommended to undertake extensive studies based on combining various remotely sensed data sets and ground data for snow cover monitoring and modelling. The synergism and complementarity of remote sensing data sets is potentially useful for extending temporal series (e.g. merging radar and visible data sets to monitor snowpack ablation), for addressing important science issues (e.g. combining SAR and visible data to derive backscatter-albedo relationships and utilize SAR data to obtain snow albedo under cloudy conditions), and in addressing problems of sub-pixel resolution (e.g. merging SAR data with SSM/I data).
- 3) Because snow cover is characterized with wide geographic and high temporal variability, large amounts of ground data are required to develop and validate robust snow retrieval algorithms. While many agencies collect snow data, large amounts of data are not stored in digital form. Further, the sampling frequency (bi-weekly to monthly) often prevent their use for validating snow retrieval models. Agencies that routinely collect snow data should undertake steps to convert their existing data sets into digital form, to store future data sets into digital form, and to increase the sampling frequency, preferably to a daily interval.

- 4) To put more effort in directing remote sensing technology toward operational use, specifically by developing information systems based on GIS technologies.

### **Snow Detection**

Satellite data, in particular passive microwave (SSM/I, SSMR), and high frequency coverage AVHRR data, have shown to be extremely useful for estimating snow water equivalent and snow cover extent at regional to global scales. However, many problems are still unresolved, for example sub-pixel heterogeneity of land cover and the resulting effect on microwave brightness temperatures. The parameterization of snow properties for inclusion in GCM's is another important issue, as well as ensuring long and consistent remotely sensed data sets for climate studies.

### **Recommendations**

- 1) NOAA/NESDIS snow charts represent a unique "long term" data set to study the temporal variability of the snow cover extent at continental to global scales. Problems related to the consistency of the data set have been successfully addressed. The NOAA/NESDIS weekly snow charts and digital archive program should be pursued and the charts (existing and future) be corrected for consistency of the data set (e.g. using the Rutgers Routine).
- 2) During the next decade many satellites carrying onboard optical and microwave sensors will be launched for Earth monitoring, and new and more sophisticated snow retrieval algorithms will be developed. There should be provision for an overlap period between existing (e.g. AVHRR, SSM/I) and future data sets for ensuring compatibility of the data for long term climate studies, and for calibration and validation of snow retrieval algorithms.
- 3) Land cover and land use must be considered for retrieving snow water equivalent, snow cover extent, snow wetness, and snow albedo from remotely sensed data. Digital land use/land cover maps at a minimum resolution of 10 kms should be developed for regions with seasonal snow cover. Agencies and groups, such as the national Snow and Ice Data Center (NSIDC), the SSM/I Product Working Team (SPWT), the Canadian Climate Centre (CCC), the Institute for Space and Terrestrial Science (ISTS), and the University of Alaska in Fairbanks, should coordinate their efforts to address this important issue.
- 4) Seasonal snow cover is a very important component of the energy exchange over land and sea, yet GCM's represent the snow cover in a simplified way. The parameterization of snow parameters, which is required because of the coarse resolution of GCM's, necessarily introduces errors in GCM's outputs. It is recommended that GCM's and remote sensing specialists increase their degree of collaboration for defining acceptable levels of accuracy for snow parameters derived from remotely sensed data. This should include spatial resolution (10, 100 km, etc.), and accuracy of the value itself (5%, 10%, etc.).

### **Algorithm Development and Science Issues**

Issues related to algorithm development were brought up at the Snowwatch'85 meeting. In particular, it was recommended that snow depth and water equivalent algorithms derived from passive depth and water equivalent algorithms derived from passive microwave data be further developed and refined. Some progress has been made, for example in the

development of snow retrieval algorithms from passive microwave data applicable to mountainous terrain. However, considerable work is still required, in particular in calibrating and validating the algorithms.

### ***Recommendations***

- 1) The spatial (local to regional to continental to global) and temporal variability (seasonal and within a season) of snow cover requires further research. An improved understanding of snow pack variability will result in an improved parameterization of those important snow pack variables for the calibration and validation of snow retrieval algorithms.
- 2) Key regions of high potential sensitivity to climate change should be identified, based on observed and modelled data. Among the possible regions are: North West of North America, North East of North America, and North East of Siberia. Such sites could be used to extensively monitor the temporal and spatial evolution of the snow cover, and to develop and test snow retrieval algorithms.
- 3) Snow albedo is a key parameter in GCM's. The variation of albedo with snow depth and land cover/land use type needs further research.
- 4) Research is required to increase our understanding of snow-cloud interactions and the remote sensing of those interactions.



## Working Group #2: Statistical Procedures

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*Francis Zwiers, Rapporteur*

### **Data Homogeneity Issues**

The utility of snow and ice data for climate change detection and the monitoring of long time scale climate variability depends crucially upon the internal integrity and homogeneity of the snow, lake ice and ice cap records. Changes in homogeneity which are caused by changes in observational practices, instrumentation, instrument location, quality control procedures, interpretation methodology and archiving policy and methodology all have a deleterious effect on our ability to use the data for climate study purposes. Automation, new sensing techniques and economics put the homogeneity of many longer-term and future series of observations at risk. These necessitate the development of techniques for record merging or other transition techniques that ensure the utility of records for climate change detection. This is particularly a problem for satellite derived data because of the fast pace of technological change and the relatively short life of satellite instruments. Analysing the brightness temperature without algorithm modification would obviate concerns about changing algorithms.

A relatively constant aspect of satellite based cryospheric data is the NOAA snow cover chart which is derived manually from visual imagery. This record, which is now 25 years in length, is likely the single most important cryospheric indicator of climate change currently available. This working group can not urge the continuance of this record strongly enough.

Long term cryospheric indicators obtained at the surface, such as the date of freeze-up and break-up on lakes and rivers which are free of the effects of human development, snow course data, and solid precipitation data obtained from gauges also have the potential to be very useful indicators of regional and global climate change.

However, the utility of these data depends critically upon the quality of the metadata which has been kept at each station (about the type of gauge, its sighting and exposure, etc). The utility of gauge data also depends upon our ability to transform observations obtained from each gauge type to an accepted standard, such as the new double fence designs now under development.

### **Recommendations**

- 1) Retain and lengthen current cryospheric indices including the NOAA snow cover charts and time series of surface data such as the dates of freeze-up and break-up on lakes and rivers and the snow depth recorded on snow courses. Continue to promote gauge and snow course standardization, the WMO programme for solid-precipitation gauge intercomparison, and the development of algorithms and other guidance for the conversion of existing gauge records to a standard reference base.
- 2) Continue development of a standard, accurate precipitation gauge, such as the double fence design now under consideration, and develop algorithms for the conversion of existing gauge data to this standard. The latter are fundamentally important if surface data is to make a substantial contribution to the diagnosis of climate change and variability.

- 3) Subject existing long term cryospheric indicators to careful change point and intervention analyses to insure their suitability for use in studies of long term climate variability and change.
- 4) Reiterate *Recommendation 2* of the "Detection of Trends" section of the Snow Watch '85 recommendations (see page 7). This recommendation stated that

Surface stations records for longer periods (  $\geq 50$  years) should be examined for trends and low-frequency fluctuations. The optimal strategy will likely involve the choice of a network of (  $< 100$  ) "benchmark" stations having homogeneous records minimally affected by urbanization and by changes in locations, instrumentation, etc. Emphasis should be placed on climatic "key regions" in the Snow Transition Zone, which undergoes large seasonal migration. We recommend that the WMO appoint a working group of data experts to address the selection of stations (regions) and appropriate parameters to observe snow depth, deviation, depth threshold, etc.).

- 5) Immediately initiate studies to determine the statistical methodology which will be used to integrate the current NOAA snow cover charts with products which are anticipated from passive and active microwave remote sensing of the earth's surface. These studies should also determine the amount of overlap required to ensure that the conversion algorithms which are obtained are accurately parameterized and will produce a single homogeneous record from the visual and microwave based charts.

### **Statistical Methodology Issues**

Recommendations 2)-5) above all touch upon the issue of the spatial and temporal homogeneity of data which are obtained from a variety of sources and instruments. The statistical methods which are used to detect changes in the characteristics of data and to subsequently adjust the data are developing and becoming more sophisticated. However, these methods are far from mature and are not as well developed or powerful as we would like. This state of affairs prevails both in the time and space domains. Therefore, in view of the importance of this methodology, particularly for the detection problem, we make the following recommendation.

#### ***Recommendation***

Encourage and foster the continued development of statistical methodology for a) the detection of changes in data homogeneity over time and space; and b) for the determination of appropriate adjustments to "old" measurements to make them compatible with current instrumentation and observational practices.

It was noted during the workshop that snow depth is frequently underestimated as a consequence of the reporting of *trace* amounts of snow. Current practice with regard to these reports is to set the corresponding snow depth to zero. However, statistical methods which are designed to utilize censored data may be able to extract additional information from these reports.

## ***Recommendation***

Reports of trace amounts of snow should be regarded as *censored* observations. Research should be undertaken to determine the feasibility of using statistical techniques which are designed to handle censored data to improve estimates of mean snow accumulations. It may be possible to use these techniques both in the context of constructing regional estimates of snow amount and extent of cover at a particular time and in the context of constructing local estimates of time averaged snow amount.

Reliable estimates of the spatial and temporal variability of liquid and solid precipitation are necessary for a variety of reasons. A few examples include: (a) detection studies in which we need estimates of spatial and temporal variability to place current variations in space and time in a correct historical context; (b) diagnostic studies of the mechanisms of the climate system; and (c) studies of the existing and planned observing networks using the methodology of spatial statistics to determine the spatial and temporal *representativeness* of station data. Current characterizations of spatial and temporal variability frequently rely upon EOF analyses to provide the necessary correlation structure estimates. However, EOF analysis is not parsimonious and consequently, does not always produce estimates which are reliable for all the purposes outlined above. For example, studies of the spatial and temporal representativeness of station data require detailed descriptions of the local spatio-temporal correlation structure reflecting the micro-climates of the regions containing the stations. The development of these local correlation function models will require the collaboration of climatologists and spatial statisticians. The replication of such an effort at a number of geographically scattered locations with different local micro-climates may ultimately provide the insight which is necessary to improve upon current practice with regard to the analysis of the structure of spatial and temporal variation.

## ***Recommendations***

- 1) Collaborative research should be initiated to find parsimonious ways to represent the *joint* spatio-temporal correlation structure of physical processes such as solid and liquid precipitation.
- 2) A collaborative demonstration project involving both spatial statisticians and climatologists be initiated to assess the spatial representativeness of solid and liquid precipitation station data at several locations which exhibit a variety of micro-climates.

Because there are only a limited number of long snow and ice records and because the satellite based global scale record is only now 25 years in length, it will be necessary to base climate change detection efforts primarily on one-number indices. An example of such an index is the snow cover index which is obtained from the NOAA snow charts. Another example is any index which is obtained from a "finger-print" (essentially a weighted average of point data in which the weights are chosen in an optimal way according to prior information about climate change obtained from climate simulations). Our assessment is that the best finger prints which can currently be derived will be based upon the results of equilibrium doubled CO<sub>2</sub> experiments. We recognize, however, that these indices are not likely to provide indices of climate variation which reflect the anthropogenically induced climate signal in an optimal way.

## **Who Defines the Null Hypothesis (for detection and other issues) - and how?**

We recognized that, while climatologists as a group may not have full facility with regard to statistical methodology, good physical insight is required to develop appropriate hypotheses and to indicate which indices and measures are important. Working closely with statisticians appears warranted in addressing many issues such as intercorrelations, the development of a most appropriate snowcover index, and problems of scale and data aliasing.

Having said that, we also recognize that *data dredging* is an issue which climatologists in general should be more aware of. Data dredging is the circular process of using "exploratory" techniques to study a data set, using the information so derived to develop hypotheses and consequently testing these hypotheses using information from the same (or a closely related) data set.

### **What do we perceive to be the impediments to change detection and the characterization of snow and ice variation?**

*Impediments to the characterization of spatial variation:* As discussed above, current methods based on EOF analysis are not parsimonious enough to give fully reliable estimates of the primary modes of spatial variation.

*Impediments to the characterization of temporal variation::* The main impediment is the relatively short length of the currently available data sets. A secondary problem is the artificial variability which may be induced by sometimes frequent and sometimes undocumented instrumentation and analysis changes.

### ***Recommendation***

As full a characterization as possible of the spatial and temporal variability of snow and ice should be developed from physical reasoning to supplement information from data. Such a derivation is needed to assess, a priori, whether change detection is possible with currently available snow and ice data.

# **Working Group #3: Lake Ice Cover as a Climatic Indicator**

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*R.G. Barry, Rapporteur*

## **Climatological Aspects**

- 1) Freeze-up/break-up of freshwater bodies is well correlated with local air temperature values during the transition seasons. The relationships appear to be stronger in continental/drier climates. Snow depth (white ice) has a limited effect on total ice thickness in cold climates.
- 2) A  $\pm 1^{\circ}\text{C}$  temperature fluctuation corresponds approximately to a 3-6 day range in the date(s) of spring break-up/fall freeze-up, respectively.
- 3) Lake ice records provide a valuable supplement to temperature observations. They are a simple integrator of transition-season climate and in northern areas with a sparse station network their records could be used to fill in data gaps for the detection of climatic (temperature) trends.
- 4) It is preferable to identify the break-up and freeze-up dates separately, rather than solely the duration of the ice-free season.
- 5) The record of freeze-up date is usually less 'noisy' than break-up date and is generally better correlated with air temperature. Break-up date is affected by other factors such as snow depth, cloudiness, wind velocity.

## **Criteria for the Use of Ice Conditions as a Climatic Indicator**

- 1) Lake ice indices should primarily refer to "Complete freeze over" and "Total clearance". These reduce observer subjectivity and facilitate identification by remote sensing methods.
- 2) Definitions of ice conditions are provided by "MANICE" (Atmospheric Environment Service. Environment Canada, May 1989).
- 3) There are no consistent definitions of ice-covered/open water for use in remote sensing mapping of lakes. Clouds present a problem for the visible/IR wavelength range; passive microwave data suffer from the current low resolution of the sensors; radar data have high-resolution but data processing requirements and the large data volume cause problems.
- 4) Ice thicknesses vary spatially (away from shore). The influence of snow depth and ice thickness on break-up date requires additional study.
- 5) The effect of lake area and depth on freeze-up/break-up needs further investigation. Lakes  $< 3\text{m}$  depth should not be considered; the question of an upper limit to the depth of suitable lakes for monitoring is not known.



## **Data Considerations**

- 1) Several northern countries have long records of ice formation/disappearance. These have been comparatively neglected as a climatic indicator in part due to the diverse interests of the collecting agencies.
- 2) There is no global inventory or archive of data on freshwater ice formation/disappearance. The amount of data held by national agencies and the media on which they are stored, are unknown.
- 3) Most lake ice data are operational in character and there has been limited quality control.
- 4) There is some potential for historical records of freshwater ice conditions in newspapers (Finland) and archives (Canada, Russia, China and Japan).

## **Recommendations**

- 1) A directory of freshwater ice data archives should be prepared. World Data Center-A for Glaciology may be a suitable organization to undertake this (contingent upon necessary resources being available).
- 2) In view of the modest data volumes, a single archive of freshwater ice conditions could be established at an existing data center.
- 3) WMO should determine whether the specifications for ice conditions in MANICE are adequate and, if so, formally adopt them for international use.
- 4) Criteria for remote sensing mapping of freshwater ice conditions should be developed. WMO is encouraged to identify individuals/groups with appropriate expertise through the Working Group on Sea Ice and Climate.
- 5) Additional research is needed on the physical controls of lake ice formation/disappearance. Organizations such as the National Hydrology Research Institute of Canada are encouraged to investigate the roles of lake morphology, meteorological controls (especially the energy balance) and snow/ice thickness.
- 6) National agencies should take steps to digitize any analog records of ice conditions (for example, records for 1985-present in Canada; the status of records in the former Soviet Union needs investigation).
- 7) The loss of manned stations in northern latitudes will increase the likely value of remote sensing of lake ice cover. A hemispheric-scale analysis of 'representative' lakes should be conducted using available satellite data:
  - (a) Defence Meteorological Satellite Program Operational Line Scan System and NOAA Advanced Very High Resolution Radiometer visible and IR data
  - (b) Scanning Multi-frequency Microwave Radiometer and Special Sensor Microwave Imager passive microwave data.

## Working Group #4: Data Management

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*R.G. Barry, Rapporteur*

### Progress since 1985

The data-related recommendations of Snow Watch '85 were reviewed. Several groups identified a general need:

- (a) to integrate conventional surface and remotely-sensed data to augment the spatio-temporal coverage of station records and provide control or remotely-sensed data, and
- (b) to develop data sets for daily time interval and over 50 – 100 year time intervals, where possible.

More specific recommendations on data bases were:

- (i) to urge WMO to standardize the observation and reporting of snow data and to make their global exchange mandatory;
- (ii) to extract snow depth reports from the GTS data stream on a routine basis;
- (iii) to seek the reporting of information on land surface type with snow cover data;
- (iv) to ensure the continuation of the weekly NOAA/NESIDS snow charts and digital archives; and to check the quality and homogeneity of those records.

Progress has been made in the following areas:

- (1) Quality analyses of the NOAA snow charts have been carried out, resulting in some corrections of identified biases.
- (2) Daily data sets are beginning to be assembled.
- (3) Some international exchange of snow data from China and the F.S.U. is underway.
- (4) Archives of surface and remote sensing data for climate studies are being developed systematically by NOAA and by NSIDC/WDC-A for Glaciology.
- (5) The combination of recalibrated SMMR products with SSM/I products is underway.
- (6) CAC/NOAA examined snow depth data in the GTS stream for a test period but the quality was not considered adequate to develop snowfall/depth data sets. [If there were a strong community demand for these data, they could be stripped out of the GTS stream.]

## Recommended Actions

### *Conventional Data*

There is an urgent need for various high-priority data sets to be digitized and transferred to WDCs or national centers for distribution. The following were identified:

- (i) snow course data for Canada (*ca.* 1300 stations). The Canadian Climate Center should digitize these data, extract meta-data and provide the data to WDCs;
- (ii) snow course records in the Former Soviet Union (*ca.* 2,200 stations). Digitization of these has been initiated by Professor A.N. Krenke, Institute of Geography, Moscow. These will eventually be archived at WDC-A for Glaciology through NOAA/ESDIM support;
- (iii) daily station data for the F.S.U. (*ca.* 3500 master state records). Data for about 250 first-order stations covering records through 1984 have been exchanged under the US/USSR Bilateral Agreement on Environmental Protection (WG VIII). Updating and expansion of this coverage should be sought; financial support for digitizing in the F.S.U. is an urgent priority. Snow data for central Asia held by SANIGMI, Tashkent, should receive separate attention following the break of the F.S.U.;
- (iv) lake ice data (dates of freeze-up, break-up and thickness) and related nearby climatic observations should be inventoried and exchanged internationally. It is recommended that the Canadian Climate Center provide a lead in this activity, supported by a request to national representatives for assistance from WMO (WCP).

### *Remote Sensing*

Daily passive microwave data from the DMSP SSM/I sensors, collected since mid-1987, are available from various sources for operational and retrospective users. Snow cover products proposed by the SSM/I Products Working Team and being implemented at NSIDC should meet the needs for a climatological snow data set. Further work on regional and/or temporally-varying algorithms will be necessary to allow for the effects of land cover, snow grain size, and freeze-thaw events.

The question of whether, or how best, to merge conventional and remote sensing snow observations has received little attention.

### *Observing Practices*

There is a large number of gauges and measurement procedures in use by national agencies for snowfall and snow cover. An international program is underway to intercompare and correct precipitation measurements (including solid precipitation). Similar attention needs to be given to the recording of snow depth on the ground and other snow pack properties.

### *Archiving Procedures*

The development of standard procedures for reporting meta-data relating to snow data sets should be encouraged. (Information should include station location, land surface type, gauge type, observation/processing procedures, etc.).

Data sets and archival procedures employed by national and other data centers should be reviewed and coordinated with other agencies (ICSU, IAHS, WDCs and CODATA, IGBP).

The existence of supplementary data (collected by state and local agencies as well as private sector or research organizations) should be communicated to NSIDC, or other data centers, so that efforts can be made to ensure their archival.

## Working Group #5: Modelling

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*John Walsh, Rapporteur*

The Working Group on Modelling focused its discussion on (1) progress that has been achieved since 1985 with regard to the modelling recommendations made by participants in Snow Watch '85, and (2) recommendations concerning the most pressing needs today in the modelling of snow cover and its interactions with global climate.

### Progress since Snow Watch '85

Snow Watch '85 called for the development of highly detailed, physically-based snow models, together with methods for aggregating subgrid-scale information to the grid scale (200-400 km) and for disaggregating grid-scale averages to smaller scales. With regard to the development of physically-based formulations for global climate models, significant progress has been made through CLASS, the Canadian Land Surface Scheme, recently developed at the Canadian Climate Center. Unlike older schemes that use the force-restore method for the soil thermal regime and a "bucket" approach for the soil moisture regime, CLASS incorporates three soil layers with physically-based calculations of heat and moisture across the layer boundaries. Snow-covered and snow-free areas are treated separately, and snow cover is modeled as a discrete "soil" layer. CLASS is the first land surface scheme to distinguish the thermal regime of the snow from that of the soil. Moisture is variable in the three soil layers, in the snow mass, and in the canopy interception store. The surface infiltration rate is calculated using a simplified theoretical analysis allowing for surface ponding. Because of the use of multiple soil layers, freezing and thawing can proceed layer by layer. In recent model-data comparisons performed at CCC, the discrepancies between the simulated snow depths and those of the Rand Climatology are smaller with CLASS than with the old ("standard" GCM) land surface scheme. As will be noted in the following section, the primary needs of CLASS with regard to snow pertain to subgrid-scale variability. Specifically, parameterizations are needed for (a) fractional snow cover as a function of snow mass for different terrain types, and (b) snow masking characteristics of vegetation as a function of snow depth. The required compilations of data include gridded representations of surface elevation and its variability over grid-cell areas, the prevailing orientation of topographic features, vegetation type, surface type (especially lake fraction), and soil type. Although such datasets are prerequisites for progress in the modelling of snow-climate interactions, the data compilations at the required resolution (Section 2.1) are either not underway or are in the early phases. Thus, with regard to the recommendations of Snow Watch '85, progress in aggregation/disaggregation has lagged the recent progress in physically-based surface model development for GCMs.

A prominent recommendation of Snow Watch '85 was the development of an observationally-based surface albedo climatology for validation of models. Work in quantifying surface albedo variations during summer over a limited geographical domain has been performed at the University of Colorado (Serreze et al., this volume), although the albedo variations are for the Arctic Ocean rather than for land areas. Nevertheless, this effort has produced weekly albedoes for ten summer seasons. The need remains for a baseline dataset relating albedo to variations in snow cover, vegetation type and leaf area index (Section 2.2).



With regard to model applications, Snow Watch '85 recommended sensitivity studies of the atmospheric response to snow cover, with particular attention to the snow-albedo-temperature feedback. This feedback was the subject of a recent model intercomparison (Cess et al., 1991, Science) in which the feedback was evaluated for 17 global climate models. A wide spectrum of sensitivities emerged, ranging from a few negative values to positive values that varied by more than a factor of two about the mean. The magnitude and/or sign of the snow albedo feedback also depended on cloudiness in most of the model experiments. In a recent experiment with the GISS GCM, Cohen and Rind (1991, J. Climate) found that the snow-albedo-temperature feedback can be largely offset by compensating changes in the other components of the surface energy budget. These studies indicate that large uncertainties remain in our estimates of the sensitivity of climate and weather to snow cover. The uncertainties are intertwined with differences in the various models' formulations of snow cover and land surface physics, and further experimentation with state-of-the-art formulations is clearly required.

The Proceedings of Snow Watch '85 contain a recommendation to evaluate weather forecasts produced by models containing interactive snow cover. An example of this kind of verification exercise was presented by Peterson and Hoke (1989, Weather and Forecasting), who found that a 48-hour NMC model forecast of temperature and precipitation type was improved drastically by a more realistic initialization of the continental snow cover. To the best of our knowledge, similar studies have not been performed at other prediction centers. Recommendations pertaining to numerical weather prediction models are contained in Section 2.4.

Finally, the verification of the snow cover simulated by several global circulation models is described briefly in the IPCC Scientific Assessment: Climate Change (1990, p. 112-113). This assessment indicated that "several models achieve a broadly realistic simulation of snow cover," but that "there are significant errors in the snow cover on regional scales in all models" (IPCC, p. 113), e.g., eastern Asia. The IPCC assessment also notes that "detailed assessments of the simulations, especially for seasons other than winter, are hindered by the different forms of the model data (mostly seasonal mean liquid water content) and the observed data (either frequency of snow cover or maps of depth at ends of months)."

## **Recommendations**

While a number of specific recommendations emerged from the modelling subgroup's discussion, a general or "over-arching" recommendation pertains to the interplay between models and data. There is a need for closer interaction between the modelling and observational communities. As suggested by the IPCC's comment on the difficulty of validating models with data, this interaction is needed in order to ensure compatibility between modelers' data needs and the available data. In addition, the problems of scaling or aggregating/disaggregating the snow-related information for a particular grid cell will likely pervade the snow modelling activities of the next decade. At this early stage of the investigation of scaling strategies, there is a need both for model sensitivity studies and for process studies to identify the key scales for data aggregation. Process studies and field programs for the acquisition of snow data will thus need to be designed with an eye toward the data requirements of modelers. A winter component of BOREAS, for example, should address the problem of scaling local measurements and grid-square estimates of quantities affected by snow. Alternatively, the scaling strategies that are employed by modelers may need to be shaped by the feasibility of the associated observational programs.

On the basis of the above discussion, the questions posed by the workshop organizers, and considerations of feasibility, the following specific recommendations were assigned highest priority by the working group on modeling.

### ***Datasets describing key surface properties at the subgrid scale***

In order to identify the appropriate scales for aggregation/disaggregation strategies and to test corresponding parameterizations in models, there is a need for high-resolution databases depicting key land surface properties within cells of  $\sim(10 \text{ km})^2$ :

- surface elevation and its variability
- vegetation (selected from 20-30 key types)
- soil type, texture and albedo
- fraction of bare soil
- fractional coverage of lakes and urban areas

In addition, information on the orientation of major topographic features (i.e., slopes) is required at  $\sim 1 \text{ km}$  resolution in order to parameterize the uneven distribution of daytime insolation.

A database of this kind is a prerequisite for further progress in modeling interactions between the atmosphere and the land surface. Work on some of the variables listed above is ongoing: for example, the U.S. Navy Global Elevation Data Set contains elevation and terrain data at a resolution of  $(15 \text{ km})^2$ . Several fine mesh vegetation data sets are available, although not for all snow-covered regions at the required resolution. There is a particular need for data on the areal coverage of lakes because subgrid-scale lakes are presently not included in GCMs. The data on surface elevation could be summarized by the use of probability density functions, while a GIS framework may be useful in consolidating the various other types of surface data.

### ***Snow-vegetation albedo climatology***

With regard to the modeling of snow albedo variations, a primary observational need is the documentation of albedo as a function of overlying snow mass (depth) and leaf area index (LAI), for the key vegetation types noted above which occur in mid to high latitudes. The availability of these functional dependencies would permit the parameterization of the vegetative masking effect, which is a first-order determinant of the surface albedo. The baseline data must include the maximum snow albedo for each vegetation type, i.e., the albedo under conditions of deep snow and minimum LAI. Albedo determinations would also have to be made for various other snow depths and LAI values. The uncertainties in the albedoes will need to be less than 5%.

This data requirement may be viewed as a "generic" set of albedo values for high-latitude land areas under different snow conditions. An alternative strategy, the mapping of the surface albedo for specific years, may provide useful supplementary data; however, the lack of generality could make such a strategy less advantageous in the simulation of snow-covered areas under substantially different climatic regimes or in the simulation of climatic change.

Tools for acquiring the necessary data include visible and IR sensors onboard aircraft and satellites. The high resolution permitted by aircraft flights is a distinct advantage in areas of inhomogeneous vegetation, although an extensive aircraft campaign would be required if the data are to span the ranges of LAI and snow depth. Ground "truth"

pertaining to snow depth (mass) and LAI will clearly be an essential component of such a measurement program.

### ***Global Climatologies of Snow Water Equivalent and Areal Extent***

As in the report of Snow Watch '85, there is still a need for climatology of the water equivalent of snow for use in the validation of model simulations of the present climate. The water equivalent is most likely the snow variable of greatest relevance to the snow-hydrology feedbacks in the climate system. Ongoing comparisons of satellite passive microwave determinations and surface measurements (e.g., those for the Canadian prairies) are clearly a major component of this effort, although there is a need to broaden such efforts geographically. As reliable spatial fields of the distribution of snow water equivalent become available, direct comparisons with climate model simulations should be assigned high priority.

The modelling group recognized the ongoing work being done at Rutgers University with a view to extracting a reliable snow extent climatology from satellite measurements. It was recommended that these efforts be continued, since such a data set would provide a valuable secondary source of information for model validation.

### ***Incorporation of GCM snow cover formulations into NWP models***

Atmospheric models used operationally for numerical weather prediction (NWP) typically do not change the surface characteristics in response to snow melt or snow accumulation. As noted earlier, the consequences for weather forecasts can be severe (Peterson and Hoke, 1989, Weather and Forecasting). Given that formulations (e.g., CLASS) now exist for surface snow cover and the physical processes by which it changes, there is a need to implement these formulations in weather prediction models. Such implementations would (a) permit the testing and evaluation of the formulations over timescales of the synoptic systems that are largely responsible for the temporal changes in snow cover, (b) expand the capabilities of NWP models by allowing for the effects of changes in the surface state during the forecast, and (c) permit more systematic tests of the impact of snow cover on the skill of numerical model forecasts. (b) and (c) were recommended by Snow Watch '85's working groups on modeling and snow/atmosphere interactions.

### ***Process studies of possible relevance to the modeling of snow***

The modeling group and other workshop participants identified various processes that may play significant roles in the evolution of snow cover. In most cases, these roles have not been quantified and are poorly understood. Thus, from a modeler's perspective, there is a need for process studies that focus on issues such as the following:

- the dependence of snow albedo on aging and wetness
- the effects of depth hoar on the thermal and moisture budgets of a snow layer
- the role of horizontal heat transfer between bare land surfaces and snow cover in areas of patchy snow cover
- the effects of aerosol deposition on snow cover (e.g., albedo)

- the variation of snow albedo with factors such as cloud cover and solar zenith angle
- the effects of environmental factors on grain size growth in snow (possibly useful for the development of albedo parameterizations).

Observational data from field programs will likely be necessary in order to address these issues. As noted at the start of Section 2, interaction between modelers and field researchers during the design phases of field programs would help to ensure that observational strategies are meshed optimally with the needs of modelers.

### ***Inclusion of lakes and lake ice in models***

To the best of our knowledge, no global climate models include lakes that cover less than half the area of a model grid cell. (The Great Lakes and other large water bodies are included when their dimensions are comparable to or larger than the grid spacing.) Yet the large numbers of small lakes in many northern land areas represent substantial fractions of the surface area in regions such as northern Canada, Scandinavia and Siberia. Because the freeze-melt cycles can affect the surface exchanges over grid-cell areas and perhaps over larger regions, the inclusion of lakes and lake ice in GCMs merits high priority. The patchy distribution of northern lakes introduces parameterizational questions into GCM formulations of lakes and lake ice, in much the same way that vegetation and topography introduce heterogeneities that complicate the parameterization of snow cover.

The wealth of historical data on lake freeze-up and break-up can provide a means for validation of the simulated surface thermal regime in the models. Validated models can also be used to estimate the impact of changing climatic forcing (e.g., greenhouse gas concentrations) on high-latitude lakes.



# Workshop on Cryospheric Data Rescue and Access

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## 1. *Background: NOAA Earth System and Data Information Management Program*

The National Oceanic and Atmospheric Administration (NOAA) has instituted a program of Earth System Data and Information Management (ESDIM) to coordinate data and information management activities on an agency-wide basis. The specific objectives of the program are to:

- Build a top-level consensus within NOAA on data management issues, and to formulate a vision of the agency's data and information management strategy for the 1990s and beyond.
- Rescue critical NOAA environmental data currently at risk of being lost.
- Improve access to NOAA environmental data and information for scientists and administrators.
- Modernize and interconnect environmental data systems throughout NOAA to increase their capability and responsiveness.
- Assist in developing standards for data documentation, data quality, and network connectivity.
- Provide agency-wide guidelines on developing policies related to environmental data management.

The first objective is addressed in the ESDIM management plan and involves the construction of a strategic approach to data management and information that can be applied agency-wide. The Cryospheric Data Rescue and Access workshop highlights the second and third objectives, with emphasis placed on objective two--data rescue.

The early focus of the ESDIM implementation plan is on data rescue, with the cryosphere being one of the first topics to be addressed. Data rescue in this context refers to "saving" data sets that are critical for scientific research. This may involve copying data from existing magnetic tapes to new tapes or to other media; transcribing disintegrating historical paper records to digital, analog, or micro-form; or the compilation of new data sets from highly varied original sources with different media types. The



concept of data rescue, in the context of this workshop, was not limited to existing cryospheric surface data records, but was extended to include "indirect sources" such as reprocessing satellite data not previously considered in a cryospheric context, and to the consideration of potential future data sources that may not automatically be included in the cryospheric data stream. Furthermore, cryospheric data rescue was also taken to include rescuing ancillary data sets that are essential for cryospheric applications. As a final note, ESDIM's objectives specify the rescue of critical NOAA environmental data sets, while our workshop considered cryospheric data from all sources.

Data access, on the other hand, focuses on providing, or enhancing, the ability of researchers to access existing or rescued data sources. Although the primary objective of the workshop was data rescue, some degree of overlap occurred in the discussion of "rescuing" data and making them accessible to the user community.

## **2. *Cryospheric Data and Applications***

The cryosphere, representing the solid phase of the hydrosphere, occupies a unique place in the global water cycle. Ice in the atmosphere plays a vital role in the precipitation process, and at the surface of the oceans, it drastically modifies the ocean-atmospheric exchanges of heat and momentum. When ice occurs on land, it represents a major source of fresh water for societal use, and acts as a significant agent for geomorphic activity. Past changes in climate have led to variations in the extent of the cryosphere, and the effects of this can still be seen in the geomorphic and isostatic history of large parts of the middle and high latitudes. Changes in the extent and distribution of the cryosphere itself have a positive feedback on climate, and thus the magnitudes of past climatic changes are likely to be linked, in part, to the expansion and contraction of the snow and ice cover. Similarly, the Global Climate Model (GCM) predictions of future global warming through enhanced "greenhouse" effects are strongly dependent on the effectiveness of the cryospheric response.

The cryosphere interacts with the Earth System in a variety of ways--the most important of which is its interaction with climate. While climate controls on the cryosphere are readily apparent (in general outline, if not in detail), what may be less obvious is the way in which changes in the cryosphere can feed back to influence the climate system. The most notable of these feedbacks, and the one common to all parts of the cryosphere, is the positive snow/ice-albedo feedback. The high albedo of snow and ice, relative to most natural surfaces, means that any change in climate which changes the spatial or temporal distribution of the cryosphere will be enhanced by the cryospheric response. Undoubtedly, such a feedback came into play during the growth of the continental ice sheets, and in the present there is some suggestion that a similar feedback may take place on time scales of weeks-to-months due to anomalous seasonal snow cover. The positive ice-albedo feedback is also a dominant factor in the high latitude response of many GCM experiments of future global warming.

The feedbacks between the cryosphere and the rest of the Earth System have led to the suggestion that the cryosphere is not only an important component of the system, possibly instrumental in enhancing global change, but it may also be a sensitive indicator

of such global change processes. The attention focused on the cryosphere has been prompted by the important role that snow and ice plays in GCM climate change experiments, together with the recognition that the cryosphere has undergone large variations in the past that correspond to periods of extensive changes in global climate. It is important to note, however, that many cryospheric processes are not well understood and are very poorly simulated by current numerical climate models. Cryospheric data are thus essential for validating Earth Systems Models and for improving model physics, as well as for monitoring environmental change and variability. These two (overlapping) requirements are central components of the U.S. Global Change program, and they formed the primary themes for the data rescue discussions in the present workshop.

The cryosphere is an appropriate target for ESDIM's early data rescue efforts not only because of the apparent importance of these data, but also because of the nature of the data sets themselves. Cryospheric data encompass a wide range of parameters including, for example, permafrost depth and distribution; snowcover extent, depth, and water equivalent; sea ice type, thickness, and distribution; and glacier/ice sheet thickness, velocity, mass balance; and chemistry and vertical structure from ice cores. These data have operational applications in hydrology, engineering, shipping/fisheries, and off-shore development. Scientific applications, as noted above, tend to focus on cryosphere-climate interactions, but again this implies a wide range of possibilities that would include snowcover-water resource applications, atmosphere-sea ice-ocean interactions, present-day ice sheet dynamics, and paleoclimatic reconstructions.

As well as this range of possible data sets and applications, several other factors complicate decision making with regard to cryospheric data management. For example, data sets are not application specific, and where time-series of sea ice concentration data may be of use to scientists interested in modeling ice dynamics for ice forecasting, they may also be of interest to atmospheric scientists concerned with the role of sea ice in the global energy balance, and to oceanographers interested in salt fluxes and bottom water formation. The importance of any given data set will, therefore, vary according to the application concerned, which must obviously be reflected in the guidelines developed for prioritizing data sets (i.e., prioritizing data sets must also involve some prioritizing of scientific objectives).

There is also the potential for some of these data sets to be very large (particularly some of the satellite-derived products from passive microwave sensors and visible and IR data from polar orbiting meteorological satellites). Furthermore, cryospheric data have been collected by both governmental and non-governmental agencies from many different countries. This raises problems of data acquisition and highlights the problems encountered when integrating data collected over varying temporal and spatial scales that exist on different media with a wide range of formats. Some further indication of the nature of the problem is shown in earlier data surveys/inventories conducted by the World Data Center-A for Glaciology (e.g., Crane, 1979; Barry, 1984; Brennan and Barry, 1989).

### **3. *Workshop Objectives***

#### **3.1 *Prioritizing Data Sets***

The primary objective of the workshop was to derive a set of guidelines for data set selection that will facilitate the successful implementation of a cryospheric ESDIM program. Given the limited resources available for the project, the volume of data involved, and the wide range of data sets/applications noted above, it is essential that such guidelines be established early in the ESDIM program.

#### **3.2 *Identification of Candidate Data Sets***

A second objective was to identify several high-priority data sets that satisfy the criteria laid out as a result of objective one. The purpose was to begin the process of data set selection by identifying several data sets that could be used to refine the ESDIM implementation plan and to begin the data rescue process.

#### **3.3 *Data Access***

A further workshop objective was to derive guidelines for data access, both for current and "rescued" data. This did not constitute a major focus of the workshop, except for data access questions that were identified as being of particular importance for cryospheric data.

### **4. *Priorities for Data Rescue***

The results of the workshop discussions are presented under four headings:

**Demands for cryospheric data.** Priority should be given to data sets for which there is a high demand or that are important for a critical research goal. As resources are limited and as we can not anticipate all future demands for data, priority is given to research areas considered to be important today. These areas all fall within the current U.S. Global Change program: data sets identified as being important for the validation of Global Climate or Earth Systems Models, for system monitoring, and for process studies. Specific cryospheric parameters are assessed in the light of these three areas of application, and each is prioritized in terms of data coverage, duration, and frequency in the accompanying tables.

Although not considered specifically in this workshop, it should be noted that there are important linkages between the cryosphere and sea level (which is a concern over time-scales of 100 years), and between the cryosphere and hydrology (important in terms of future water resources and the timing of runoff). Both of these are important research questions that may require data not discussed here.

**Guidelines for prioritizing data sets.** A set of guidelines is presented for assessing the relative importance of prospective data sets and their priority in the data rescue effort.

**High priority data sets.** Two groups of data sets are identified--one group having high priority and worthy of immediate attention, and one group that should be considered in the data rescue effort, but having a lower priority or requiring more information before priority can be assessed.

**General recommendations.** A set of recommendations is presented for the near-term implementation of the cryospheric data rescue effort.

#### **4.1 Demands for Cryospheric Data**

As noted above, the workshop discussion focused on sea ice, snow cover, and glaciers/ice sheets, emphasizing data sets useful for parameterizing or validating large-scale GCMs or Earth Systems Models, data sets that could be used for monitoring climate change and variability, and data for developing or validating empirical or numerical models of system processes. It is also recognized that data are required to develop or validate remotely sensed products and to support specific regional programs (e.g., Global Energy and Water Experiment (GEWEX), Arctic Climate System Study (ACSYS)).

##### **4.1.1 Validation of Earth System Models**

**Sea Ice.** The results from the present generation of global climate models suggest that the model climate and its sensitivity to CO<sub>2</sub> induced climate change are greatly affected by the cryosphere, and particularly by the distribution of sea ice<sup>1</sup>. This is exemplified by the fact that most GCMs show their greatest CO<sub>2</sub> warming at high latitudes, and at least one study has shown that about one third of the temperature feedback is due to albedo changes at high latitudes.

The primary importance of the ice cover is the way in which it modifies the regional energy balance. In this regard, the most important parameters are ice concentration and extent (i.e., the largest single change in the model climate is the change from open water to ice covered ocean), which are controlled by the surface temperature and ice dynamics (determined by the local windfield, ocean currents, and the internal ice stress). While the change between ice and open water affects several elements of the surface energy budget the dominant term is the surface albedo, and climate modelers frequently point to the need for surface albedo fields at GCM grid-cell resolution. The other major factor that controls the surface energy balance is radiative transfer in the atmosphere, primarily controlled by the cloud cover. The winter cloud cover is dominated by large-scale advective processes, while the summer cloud cover is more responsive to local conditions (open water and the boundary layer structure). Given this, we can establish that the highest priority data sets will be those that verify a model's ice cover and concentration and the model's polar cloud cover. Although most of the present generation of climate models include only a thermodynamic ice model, ice dynamics will

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<sup>1</sup> We should note that this does not reduce the importance of snow cover, it merely reflects the fact that very little attention has been paid to the role of the snow cover in current GCMs.



be added in the near future, and data sets relevant to ice dynamics will also be important (e.g., ice motion, surface wind).

From the Earth System Model perspective, therefore, the highest priority data sets are sea ice extent, concentration, and thickness for the large-scale energy budget, and sea ice concentration, surface conditions (leads, fractures, and melt puddles), and boundary layer structure for polar clouds. When fully coupled ocean models are included (and ice dynamics are a routine part of the model) other parameters will be needed for model verification (e.g., ice production rates, with their associated salt fluxes, ice motion, and the ice export flux).

For Earth Systems Models, highest priority will also go to those data sets that are global in coverage and have sufficient duration to derive climatological statistics (mean and variance). Monthly data are adequate for most purposes, but the original data must be collected at sufficient temporal resolution to derive representative monthly averages.

**Snow Cover.** State-of-the-art Earth Systems Models predict the fractional snow cover, snow depth and water equivalent, and layer temperature. As noted above, little work has been done on the sensitivity of climate models to snow cover, the importance of which appears to be due to its albedo (and the related masking effect on the vegetation canopy), and to its effect on the surface hydrology. Snow cover data have been generally regarded by modelers as having a lower priority than sea ice; the required data sets for model validation are snow cover extent, snow water equivalent, and snow depth (for land surface models that allow vegetation masking by snow accumulation). Again the data should be global and of sufficient duration (>10 years) to derive climatological statistics. Both conventional and remotely sensed snow data are required.

A possible future development could be a move towards the use of statistical methods to derive the spatial variability of the snow cover parameters within a grid cell. This would require the collection of data at a higher spatial resolution than the surface grid of the model (i.e., 1/2 degree or better) in order to derive the spatial relationships for use in the model.

**Glaciers and Ice Sheets.** These are currently regarded as fixed and permanent boundary conditions of GCMs. Little effort has been made to develop these components; the only data requirements at present are for an accurate map (digital) of the fractional ice-covered area within a grid cell, and for the topography of the larger ice sheets (that cover multiple GCM grid-cells).

**Permafrost and Ground Ice.** Interactions and feed-backs between permafrost and climate (mainly via greenhouse reinforcing as a consequence of melting organic matter) involve very long time scales, but alterations due to continued or even accelerated warming could be dramatic over vast areas (affecting land subsidence, coastal erosion, drainage patterns, slope instability, etc.). Borehole temperatures in ice-supersaturated and, hence, impermeable permafrost provide extremely clear signals of secular warming trends and recently accelerated temperature increases. Commercial boreholes have been drilled in connection with Arctic oil exploration, but the data are not always available for long-term temperature measurements. A few research boreholes of limited depth have



recently been drilled at high altitude/low latitude sites in Canada, China, and the European Alps.

Although permafrost was not considered in this workshop, these data should be included in a data rescue/access program. Systematic intercomparison and long-term monitoring programs are presently being planned by the International Permafrost Association. It is also worth noting that ground ice is predicted by the more advanced surface models (that predict the temperature in the top 2m of the soil). Seasonal freeze-thaw patterns may be important in the context of predicting changes in trace gas fluxes, while maps of seasonal freeze-thaw patterns and the distribution of permafrost would be a useful model validation data set in this context.

#### **4.1.2 System Monitoring and Change Detection**

Cryospheric data can be used for Earth System monitoring in several ways: the cryospheric data may be used alone (e.g., measuring the areal extent of sea ice or snow cover over time); they may be used as an integrator of various climate parameters (e.g., snowfall rates and duration represent an integrated effect of storm tracks and cyclone characteristics); and they may be used to support trends noted in other data sets (e.g., changes in glacier mass balance may be used to support trends noted in regional meteorological data).

For monitoring purposes the most important consideration is the length of record of the data set and its internal consistency. As with any meteorological data set these require a length of record sufficient to extract long-term trends from the short-term variability, and they require sufficient metadata to assess the data-set history and to separate natural changes from artifacts introduced by the data collection process. While global data are less critical than they are for model validation, data sets have to be available from enough regions to demonstrate that observed changes are globally significant.

**Sea Ice.** The most useful data sets for monitoring purposes are ice extent and concentration. It should be noted, though, that while regional variations are large, the interannual variability of the total ice extent is relatively small. It is possible that ice extent does not truly reflect the interannual variability in ice production, and ice thickness measurements may be a more useful acquisition. Little high quality thickness data are currently available, and ice thickness represents the highest priority sea ice data set.

High quality ice extent and concentration data have been available since the launch of multichannel microwave radiometers in the late 1970s. In terms of data rescue, therefore, the priority for ice concentration data is for data sets that cover regions for which surface data have been lacking in the past (e.g., the Eastern Arctic). These data would be useful for satellite algorithm validation, and for high-quality data sets that extend the ice concentration record backwards in time in the pre-satellite era.

**Snow Cover.** The most obvious snow cover data sets for climate monitoring purposes are snow extent, fractional cover, snow water equivalent, and snow depth. Also

of interest is snowfall (rate and duration), as well as other derived quantities such as the number of days per year with snow depth greater than a given threshold, the beginning and end of continuous snow cover, the date of the first snowfall, and the date of snow melt. These data sets could be regional (rather than global in extent), and are more likely to be useful when examined in conjunction with other climate indicators.

**Glaciers and Ice Sheets.** Mass changes of mountain glaciers are an important source of information on shorter-term climate change and variability. The mass changes of smaller ice caps and mountain glaciers, as well as borehole temperatures in cold firn areas, provide clear signals of the changing energy balance at the earth surface. The main advantage of the glacier signal is the possibility of quantitatively assessing rates of change and ranges of natural variability back in time. Glacier mass balance, however, is strongly dependent on local topography and microclimate, and changes in individual glaciers may not represent changes in the regional climate.

Data, then, are required from a large number of glaciers for which the relationship between mass balance and regional climate are known. Consequently, the highest priority glacier mass balance data sets are those that expand upon existing regional networks, and those for which extensive climatological studies have already been undertaken. Worldwide glacier observations started a century ago as an activity of the International Commission on Snow and Ice and continues today as the World Glacier Monitoring Service (WGMS). WGMS, headed by W. Haeberli at ETH, Zurich, is supported by the International Association of Hydrological Sciences, the United Nations Environmental Program Global Environmental Monitoring System and UNESCO's International Hydrological Program. But global coverage is still limited and the most extensive network is currently found in the European Alps<sup>2</sup>. There is a major concern that long-term data sets from the former Soviet Union may be in danger of being lost, and that ongoing monitoring programs may be reduced.

Changes in mass balance of both mountain glaciers and larger ice caps may also have a significant impact on global sea level. Although large changes in the Greenland and Antarctic ice sheets are not anticipated over short time-scales (i.e., decades), even small changes have the potential for affecting sea level due to the large volume of ice involved. There continues to be the possibility that large dynamic changes may occur in the West Antarctic ice sheet over time-scales of decades to centuries in response to past climatic forcing.

The most useful data sets for monitoring glaciers and ice sheets are those that give information about current conditions as well as changes in ice extent, ice thickness, and mass balance. Considering the size and location of the ice masses, these data usually

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<sup>2</sup> Evidence from the Alps suggests that the loss of mass involves an energy influx roughly comparable to the assumed anthropogenic greenhouse forcing, the mass loss was much greater in the last decade compared to the long-term mean, and glaciers are smaller now than at any time in the last 5,000 years (and possibly smaller than at any time in the Holocene).

consist of a combination of satellite<sup>3</sup> and field observations, which are most effective when used together. Ice cores from glaciers and ice sheets are also a valuable source of long-term climate change data (see **General Recommendation, 6**), p. 291, and long-term records of snow deposition on the ice sheet could be a useful indicator of potential climate change.

#### **4.1.3 Process Studies/Process Model Development and Verification**

Data priorities for process studies and models are less easily defined than they are for the Earth Systems Models or for climate monitoring. Virtually any data at any temporal or spatial resolution are potentially useful for empirical analysis, as inputs for process models, for model parameterization, or for model validation. In this case, the greatest priority would go to those data sets that comprise a suite of co-located measurements of several related parameters.

#### **4.2 Guidelines for Prioritizing Data Sets**

- 1) Candidate data sets that fit the Global Change research agenda as defined above have the highest priority.
  - The three research areas have equal priority (4.1.1, 4.2.2, 4.1.3).
  - Other applications should not be excluded, but they should receive a much lower priority level.
- 2) Only data sets with a certain minimum level of accompanying metadata should be considered.
  - Metadata do not have to be complete.
  - Enough information is needed to determine the exact nature of the data collected, the location and period of coverage, and to evaluate the data reliability. It should be recognized, however, that the reliability may not be immediately obvious. The evaluation of reliability often comes from the research process and through comparison with other data sets.
  - What defines a minimum level of metadata will vary from application to application.
- 3) Data in danger of being lost should get a higher priority for data rescue.
  - Priority is determined by length of time before the data set is likely to be lost.

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<sup>3</sup> Satellite data have often been used to measure snowlines (with their poor signal-to-noise ratio). We should encourage the measurement of glacier length changes, which give a better assessment of glacier mass loss at decadal and long-term time-scales.

- Higher priority should also go to data that are in no danger of being lost, but which are presently inaccessible to the user community.
- 4) Data quality is an important consideration, but it is difficult to quantify.
- Requirements for data accuracy and reliability will vary from application to application.
  - A lower quality may be more acceptable for unique data sets compared to those for which other alternatives are available.
  - The decision on an acceptable data quality (as with the decision on what constitutes sufficient metadata) will rest with the Data Center manager<sup>4</sup>.
- 5) Having satisfied the previous criteria, data sets are prioritized according to data set attributes:
- Coverage.
  - Duration.
  - Frequency.
  - Cost.

The requirements in these categories will vary for each application as indicated in Tables 1-3. There is an increasing dependence on satellite data for system monitoring and Earth Systems Models for climate change analysis; consequently, the highest priorities for data are those that match the introduction of new satellite systems and the transition between each generation of satellite systems. Equal priority will be given to the 10 year (1979-1988) Atmospheric Model Intercomparison Project (AMIP) period used for GCM intercomparisons. Similarly, high priority will also go to data sets having a spatial coverage and duration that matches other extensive data collection/analysis programs such as GEWEX, ACSYS, The International Satellite Cloud Climatology Project (ISCCP), etc.

Ultimately, decisions about which data sets to rescue will depend on the cost of the rescue operation as well as scientific importance. A cost-benefit analysis must be part of any decision making process--where costs are high and the importance of the data set relatively low, the data set would automatically receive a low priority. Low costs, on the other hand, should not automatically raise the priority of the data set. It is likely that decisions regarding costs can only be made by considering the data rescue in terms of opportunity cost (i.e., the loss of other data sets that might alternatively have been rescued). Again, this is a decision that would appear to rest with the data center

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<sup>4</sup> See, however, **General Recommendation, 5** , p. 291.

**TABLE 1. Sea Ice Data Requirements and Priorities for System Modelling and Climate Change Monitoring**

PARAMETERS	DATA CHARACTERISTICS			DATA APPLICATIONS		
	Coverage	Duration	Frequency	Monitoring	Global Models	Process Modelling
Concentration	Hemispheric/ Regional	>10 yrs	Weekly/ Monthly	1	1	1
Thickness	Regional/ Point	>10 yrs	Monthly/ Seasonal	1	1	1
Met Obs	Regional	>10 yrs	Daily	1	1	1
Sfc Radiation	Regional	>5yrs	Hours/ Daily	1	1	1
Albedo	Hemispheric/ Regional	>10 yrs	Weekly	2	1	1
Snow Depth	Regional/ Point	>5yrs	Weekly	2	2	1
Extent	Hemispheric/ Regional	>10 yrs	Weekly/ Monthly	1	1	5
Leads/Polynyi	Regional	>5yrs	Weekly	4	2	1
Ice Temperature	Hemispheric/ Regional	>5yrs	Weekly/ Monthly	3	2	2
Ice Type	Hemispheric/ Regional	>10 yrs	Weekly/ Monthly	2	5	1
Ice Motion	Hemispheric/ Regional	>5yrs	Weekly	5	2	2
Melt Ponds	Regional	>5yrs	Daily/ Weekly	3	4	2
Floe Size Distribu- tion	Hemispheric/ Regional	>5yrs	Weekly	5	5	2
Ridge Statistics	Regional	>5yrs	Weekly	5	5	2

Ranking: 1 (high) - 5 (low)

#### 4.2 Guidelines for Prioritizing Data Sets

- 1) Candidate data sets that fit the Global Change research agenda as defined above have the highest priority.
  - The three research areas have equal priority.
  - Other applications should not be excluded, but they should receive a much lower priority level.
- 2) Only data sets with a certain minimum level of accompanying metadata should be considered.



**TABLE 2. Snow Cover Data Requirements and Priorities for System Modelling and Climate Change Monitoring**

Parameter	DATA CHARACTERISTICS			DATA APPLICATIONS		
	Coverage	Duration	Frequency	Global Climate Monitoring	Earth System Models	Process Models
Snow Cover	Hemispheric/ Regional	>10 yrs	Daily/ Weekly	1	1	1
Snow Liquid Water Content	Hemispheric/ Regional	>10 yrs	Daily/ Weekly	2	1	1
Snow Depth	Hemispheric/ Regional	>10 yrs	Daily/ Weekly	1	2	2
Precipitation Amount	Hemispheric/ Regional	>10 yrs	Daily/ Weekly	2	1	1
Precipitation Type	Hemispheric/ Regional	>10 yrs	Daily/ Weekly	2	1	1
Canopy (Vegetation)	Hemispheric/ Regional	>10 yrs	Once	1	2	2
Albedo	Hemispheric/ Regional	>10 yrs	Daily/ Weekly	2	1	2
Surface Air Temperature	Hemispheric/ Regional	>10 yrs	Daily/ Weekly	5	3	1
Snowmelt (Runoff)	Hemispheric/ Regional	>10 yrs	Daily/ Weekly	5	3	1

**Ranking: 1 (high) - 5 (low)**

- Metadata do not have to be complete.
  - Enough information is needed to determine the exact nature of the data collected, the location and period of coverage, and to evaluate the data reliability. It should be recognized, however, that the reliability may not be immediately obvious. The evaluation of reliability often comes from the research process and through comparison with other data sets.
  - What defines a minimum level of metadata will vary from application to application.
- 3) Data in danger of being lost should get a higher priority for data rescue.
- Priority is determined by length of time before the data set is likely to be lost.
  - Higher priority should also go to data that are in no danger of being lost, but which are presently inaccessible to the user community.

**TABLE 3. Ice Sheet and Glacier Data Requirements and Priorities for System Modelling and Climate Change Monitoring**

	DATA CHARACTERISTICS			DATA APPLICATIONS		
	Coverage	Duration	Frequency	Global Climate Monitoring	Earth System Models	Process Models
Snow Accumulation	Greenland/ Antarctica	>10 yrs	Annual	2	2	1
Ice Extent	Global	>10 yrs	Annual	1	5	1
Ice Thickness	Global	>10 yrs	Annual	1	5	1
Ice Sheet DEM	Greenland/ Antarctica	>10 yrs	Decadal	1	5	1
Ice Velocity/Surge Activity	Global	>10 yrs	Annual	2	5	2
Iceberg Production	Greenland/ Antarctica	>10 yrs	Annual	2	5	3
Mass Balance	Global	>10 yrs	Annual	1	5	5
Land Ice Fraction	Global .5°x.5°		Long- Term Mean	5	3	5

**Ranking: 1 (high) - 5 (low)**

- 4) Data quality is an important consideration, but it is difficult to quantify.
  - Requirements for data accuracy and reliability will vary from application to application.
  - A lower quality may be more acceptable for unique data sets compared to those for which other alternatives are available.
  - The decision on an acceptable data quality (as with the decision on what constitutes sufficient metadata) will rest with the Data Center manager<sup>1</sup>.
- 5) Having satisfied the previous criteria, data sets are prioritized according to data set attributes:
  - Coverage.
  - Duration.
  - Frequency.
  - Cost.

1. See, however, **Recommendation 5**, p18.

manager, but the user community could be involved in the decision if **General Recommendation, 5**, p. 291, is implemented.

### 4.3 Candidate Data Sets

The candidate data sets are divided into two groups. The first is a group of high-priority data sets that require immediate attention. The second is a list of data that have lower priority or that require further information be obtained before their priority can be assessed.

#### 4.3.1 High Priority Data Sets

**Sea Ice and Meteorological Data from the Former Soviet Union.** Several high-priority data sets are available from drifting ice stations, coastal stations, and the "Flying Lab Program." The ice station data include surface meteorological observations, the surface radiation budget, and the snow cover on the ice (snow depth and physical properties). These data sets meet all of the criteria established above for high-priority data sets: they have application in all three research areas, there is a considerable amount of accompanying information (metadata), and there is a serious danger of these data sets being lost (conversely, there is also a window of opportunity for obtaining these data at the present time). Also, information we have obtained indicates that these are all high-quality data sets, and in terms of coverage, duration, and frequency, they cover the eastern Arctic Basin (for which we have few surface observations), are long-term, and were collected at the routine meteorological observation times. Attempts are already being made to obtain these data from the Arctic and Antarctic Research Institute (AARI), St. Petersburg. Roger Colony, University of Washington, is involved in these efforts.

Similar surface radiation budget data were collected at Arctic coastal stations. These data are of equal importance to the drifting-station data, and, at present, no attempts are being made to obtain these data. The coastal station data, therefore, have a very high priority for data rescue.

The Flying Lab Program was a series of aircraft flights into the Arctic Basin. Each flight involved landings on the ice during which surface and meteorological data were collected. The program lasted five years with approximately 300 landings on the ice per year. The data collected included snow depth, temperature, pressure, and ridging statistics. These data also satisfy the requirements for a high-priority data set, although the coverage and frequency of observation are not as high as we would wish. On the other hand, snow depth and ridging statistics are parameters for which we presently have little information, and these data would complement the drifting-station data described above. The U.S. ran a similar, but smaller, program in the western Arctic. If these data could be located (Joe Fletcher, NOAA, is a possible source of information), taken together they could form a very useful data set.

The former Soviet Union also compiled long-term basin-wide sea ice extent and concentration data. Although the U.S. already possesses high-quality data sets of this type, the Soviet data would be very useful in that they would provide surface truth for the Eastern Arctic (for which we have little information other than passive microwave

observations) and, combined with U.S. and Canadian sources for the western Arctic, this would improve on data quality in the pre-satellite era. The National Snow and Ice Data Center (NSIDC) at the University of Colorado is currently working with AARI to obtain data back to 1967, but earlier records from 1945 (and even pre-war) for the Arctic are available in map form.

Oceanographic and meteorological data were also collected to support operations on the Northern Sea Route and oceanographic data were collected for the Siberian Shelf. Little information was available to the group on these data sets, but these could be very useful for improving and validating ocean circulation models in the polar basin. Again there is a window of opportunity for obtaining these data, and immediate steps should be taken to determine their potential utility while establishing priority for data rescue.

**Snow Data in the Global Telecommunications System (GTS).** The GTS data present a potential source of long-term snow data for the Northern Hemisphere. Daily data are available for precipitation, snowfall, snow depth, Tmax and Tmin for 1976-1992 (with a possible extension back to 1967). The primary data usually include observations every 3 hours from approximately 7,500 stations world wide. The observations also include a code for present weather. These are point data but the spatial density is sufficient that this could be a useful data set for validating satellite-derived products including NOAA's weekly snow charts. Further, the period of record is long enough that these data may be useful for climate monitoring. In this respect, they have the added advantage that they are collected simultaneously with other meteorological data. These data are provisionally rated as high priority, although some preliminary analysis suggests that the data quality may be questionable. For example, the data are often recorded in different units (which is not a problem if the units are known), and there is some confusion between zero measurements and missing data. This would suggest that obtaining a usable data set will involve an extensive processing and quality checking program, rather than a simple program of data extraction.

**Glacier and Ice Sheet Data.** In general, it would be useful to have glacier data for 200-300 km drainage regions (that can be gridded as desired). These data would include glacier area, estimates of glacier volume (mass), estimates of change in volume and area with time, and statistics on the advance and retreat of glaciers (as is done for the Austrian Alps).

More specifically, Canadian and U.S. glacier data for the western mountains should be digitized and combined into a single data set. This would provide a comprehensive regional glacier data set for an area that has extensive meteorological data, and for which a considerable amount of glaciological and glacier-climate information exists. This would satisfy the criteria for a high priority, long-term, climate monitoring data set.

Every effort should also be made to ensure the preservation of data from the former Soviet Union, and to encourage the continuation of glacier monitoring programs in this region.

#### **4.3.2 Data Sets Requiring Additional Information and Lower Priority Data U.S. Navy Submarine Sonar Data (Under Ice Profiles): Ice Thickness and Water Properties.**

If submarine tracks can be located within +/- 5 miles and +/- 10 days these could provide an important data source on ice thickness distributions. Some data are in chart form, while some are on old tapes. There is also some fear that the Arctic Submarine Laboratory (where the data are currently located) may be disbanded in the near future. At present we do not have enough information on the frequency and spatial coverage of these data to determine their usefulness.

**Siberian River Runoff.** Several copies of these data are in the hands of U.S. researchers. It would be worth placing an official copy in NSIDC. (Source: St. Petersburg)

**Fractional Coverage of Land Ice.** A digital map (1/2° grid) of the fractional area of land ice would be a useful product for input to GCMs. (Source: to be compiled from Glacier Inventory Project at the World Glacier Monitoring Service)

**Russian Station Data.** 280 stations from 1984 to the present. Data include temperature, precipitation, and snow depth. The data prior to 1984 have been rescued and are archived at the National Climatic Data Center in Asheville, North Carolina. Quality control on this product has been carried out by David Robinson, Rutgers University. (Source: State Hydrometeorological Service, WDC-B, Obninsk)

**Snow Depth Data from China.** Daily snow depth data from Chinese stations from the 1950s to 1980. This data set would be useful if digitized and held in the World Data Center for general use. Data from 170 stations in the westernmost provinces are held by David Robinson at Rutgers University. (Source: Lanzhou Institute)

**SPRI-NSF Antarctic Atlas.** Raw data currently in atlas form for the Antarctic surface, bed, internal layers, etc. These data are located at the Scott Polar Research Center (SPRI). They should be digitized and made more accessible to the user community. This is a case where the issue is data access rather than rescue. (Source: SPRI)

**Russian Upper Air and Surface Observations from the Antarctic.** Data from the IGY to the present. (Source: AARI)

**Upper Air Observations from Greenland.** Data prior to the 1970s should be digitized. Digital data from 1973-present should be acquired by NSIDC. (Source: Danish Meteorological Institute)

**Canadian Snow Cover Data.** Data for 1300 stations from the 1950s to the present should be digitized. (Source: Atmospheric Environment Service, Canada)

**Satellite Data.** Many of the satellite data sets of cryospheric areas have been safely archived and are available to the scientific community at reasonable cost.



However, there are a large number of Landsat TM and MSS images with 30m and 80m resolution that were acquired under the commercialization act that are not readily available because of their prohibitive cost. Many of these images cannot be examined because they do not exist in hard copy. It would be valuable to transfer more of these data to the public domain.

Over 80,000 images are contained in the Space Shuttle Earth Observation Program (SSEOP) photography data base. This data base should be checked for useful cryospheric information.

There is a large quantity of high-resolution manned-satellite (Soyuzkarta) image data of cryospheric areas acquired by the former Soviet Union. It would be worth determining whether there will be problem maintaining this archive in Russia.

#### **4.4 General Recommendations**

- 1) Given limited resources we recommend that:
  - The majority of candidate data sets compete for funds on the basis of priority determined by the preceding guidelines.
  - A small amount of money should be available to collect and store low-cost data sets that might become available but which do not fit into these guidelines (i.e., they do not fit the priorities listed here, but the cost of collecting and storing them is minimal).
  - Some funds should be available for data sets that do not fit present scientific goals, but have high intrinsic merit (i.e., well documented, long-term, high quality, etc.). These data sets should not receive automatic high priority simply on these grounds, but provision should be made to accept proposals to rescue these data on an ad hoc basis.
- 2) Immediate efforts should be made to begin, or to assist with, the rescue of the high priority data sets listed in Section 4.3.
- 3) Immediate efforts should be made to obtain further information on the "Other Potential Data Sets" listed in Section 4.3.
- 4) If efforts are undertaken to rescue the high-priority data sets identified above, NOAA should consider a follow-up meeting in the spring of 1994 that would review the data rescue procedures and focus on identifying other high-priority data sets for the next stage of the rescue effort.
- 5) The National Snow and Ice Data Center should establish a list server on Internet that could be used periodically to update interested scientists on the status of particular data sets. The same process could be used to canvas rapidly the user community on questions of data set quality, etc.

- 6) This workshop focused almost entirely on the recent time period. We should not lose sight of the fact that the best verification of an Earth Systems model's ability to predict climate change is to run the model for past climate states. Paleoclimatic data are vital for this approach, and a separate group should address the question of data rescue for paleoclimate data sets in this context.
- 7) The group did not have the expertise to assess the needs for ground ice and permafrost data. Such expertise should be obtained for any future meetings of this cryospheric group, or a separate group should be formed to address this issue.
- 8) The order of processing for rescuing data sets should not automatically start with the oldest records and work forward (unless the oldest records are deteriorating at a rate that makes this approach necessary), nor should it start at the present and work backwards. As was noted in the criteria for data-set prioritization, the most useful data such as those that match other large-scale experiments or overlap with the introduction of new satellite systems should be processed first.
- 9) To reduce similar data rescue problems in the future, NOAA should require that all NOAA-supported investigators provide well-documented copies of all data arising out of the project to the relevant national data center. NOAA should encourage all U.S. government agencies to adopt a similar policy.
- 10) The question of data rescue, in general, and cryospheric data rescue in particular, should be put on the agenda of the annual meeting of the U.S. National Weather Service and the Canadian Atmospheric Environment Service.
- 11) Data rescue should also be placed on the agenda of the Joint U.S.-Canadian Ice Working Group.
- 12) The NOAA weekly Northern Hemisphere snow charts and the Joint Ice Center Northern and Southern Hemisphere ice charts are flagship products that currently represent the most extensive and most continuous data products for snow and ice extent. These products need to be continued and they should be produced in a consistent manner (i.e., in terms of charting techniques and data sources). Should changes be made, there should be sufficient overlap between the old and new products to allow a comparative analysis. A greater effort should be made to document these data sets and establish their temporal and spatial quality.
- 13) Few U.S. and Canadian automated meteorological stations currently collect snowfall and snow depth data. Attempts should be made to install equipment to make such measurements, at least for certain key stations. Automated stations must be capable of measuring rain and snow during all seasons.
- 14) An NSF-sponsored workshop on the U.S. Antarctic meteorological data delivery system was held in November 1987. The recommendations from that workshop have not been implemented. We would encourage NSF Office of Polar Programs to re-evaluate and to act on these earlier recommendations (Hanson and Stearns, 1988).

- 15) Several useful data sets have been compiled and released on CD-ROM; an example being the passive microwave sea ice data distributed by NSIDC. It would be useful if a general cryospheric data set could be made available in this form. This could include:
- Weekly sea ice and snow cover grids and weekly sea ice parameters from the passive microwave data.
  - Monthly and annual averages of the above three data sets.
  - Monthly and annual ice edge statistics for Antarctica (as derived by the Climate Analysis Center).
  - Combined annual global sea-level estimates derived from tide data for 100 years.
  - Monthly and annual global estimates of land air temperature (East Anglia data set).
  - Arctic ice motion data from SAR and drifting buoys. Should include the long history of T-3 locations.

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1. Unable to attend, but contributed to the report.

## ICSI Celebrates 100 Years

The International Commission on Snow and Ice (ICSI) deals with the study of snow and ice in all its naturally occurring forms, i.e., with glaciology. It was formed from the merging in 1936 of the International Commission on Glaciers (formed in 1894; one of the oldest continually active international nongovernmental scientific organizations) with the Commission on Snow of the then more recently created International Association of Scientific Hydrology (now IAHS).

ICSI deals with seasonal snow and glaciers, which are included in the hydrological cycle, as well as sea ice and with polar ice sheets. The study of deep polar ice cores yields essential information on past climates and atmospheric composition for hundreds of millennia.

The ICSI governing Bureau includes a President, currently Dr. M. Kuhn, University of Innsbruck, three Vice-Presidents, a Secretary, and four Heads of Divisions. The Divisions include:

- Seasonal Snow Cover and Avalanches
- Glaciers and Ice Sheets (including ice shelves)
- River, Lake and Sea Ice
- Ice as a Material (including ice in the atmosphere, in the ground and extra-terrestrial ice).

For particular scientific tasks, ICSI has created working groups of limited duration. For 1991-1995 the following working groups were established:

- Snow-Atmosphere Chemical Interactions
- Snow Ecology
- Glacier Mass Balance Measurements.

The main activity of ICSI is to organize symposia, especially during the IUGG General Assemblies. The proceedings of the symposia are generally published IAHS. ICSI has also encouraged the publication of several important manuals, textbooks and encyclopedias.

ICSI will celebrate its 100th anniversary with events in September 1994. The tentative schedule is as follows:

- 5-6 Sept.: Zurich. Ceremony, invited lectures, panel discussion on mass balance measurements
- 7 Sept.: Excursion to Rhone Glacier
- 8-9 Sept.: Innsbruck. Symposium on mass balance records.

For further information on the celebration, please contact: Prof. M. Kuhn, Institute of Meteorology, Innrain 52, A-6020 Innsbruck, Austria.



## GLACIOLOGICAL DATA SERIES

*Glaciological Data*, which supercedes *Glaciological Notes*, is published by the World Data Center-A for Glaciology (Snow and Ice) several times per year. It contains bibliographies, inventories, and survey reports relating to snow and ice data, specially prepared by the Center, as well as invited articles and brief, unsolicited statements on data sets, data collection and storage, methodology, and terminology in glaciology. Contributions are edited, but not refereed or copyrighted. There is a \$10 shelf stock charge for back copies.

Scientific Editor: Roger G. Barry

Technical Editor: Ann M. Brennan

The following issues have been published to date:

- GD- 1,        *Avalanches, 1977*
- GD- 2,        *Arctic Sea Ice, 1978*
- GD- 3,        *World Data Center Activities*
- GD- 4,        *Glaciological Field Stations,*
- GD- 5,        *Workshop on Snow Cover an*
- GD- 6,        *Snow Cover, 1979*
- GD- 7,        *Inventory of Snow Cover and*
- GD- 8,        *Ice Cores, 1980, Out of Print*
- GD- 9,        *Great Lakes Ice, 1980, Out of*
- GD-10,       *Glaciology in China, 1981*
- GD-11,       *Snow Watch 1980, 1981*
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- GD-17,       *Marginal Ice Zone Bibliogra*
- GD-18,       *Snow Watch '85, 1986*
- GD-19,       *Tenth Anniversary Seminar;        *7 Bibliography, 1987**
- GD-20,       *Workshop on the U.S. Antarctic Meteoro...*
- GD-21,       *Permafrost Bibliography Update, 1983-1987, 1*
- GD-22,       *Northern Libraries Colloquy, 1988*
- GD-23,       *Ice Core Update 1980-1989; Permafrost Data Work*
- GD-24,       *Passive Microwave Research; Microwave Bibliog*
- GD-25,       *Snow Watch '92. Workshop on Cryospheric Data*
- GD-26,       *Permafrost Bibliography Update, 1988-199...*

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