NORTHERN HYDROLOGY AND ITS GLOBAL ROLE

XXV NORDIC HYDROLOGICAL CONFERENCE
NORDIC ASSOCIATION FOR HYDROLOGY
REYKJAVÍK, ICELAND AUGUST 11-13, 2008

EDITORS:
ÓLI GRÉTAR BLÖNDAL SVEINSSON
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SIGURLAUG GUNNLAUGSDÓTTIR

VOLUME 1

NORDIC HYDROLOGICAL PROGRAMME
NHP REPORT NO. 50

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LIST OF CONTENTS

VOLUME 1

PREFACE

This is the fifth Nordic Hydrological Conference being hosted in Iceland, with the last one being in Akureyri in 1996. The conference is being held by the Nordic Association for Hydrology (NHF), which is an independent body aiming at promoting hydrology as a science and at increasing the understanding of hydrology and of hydrological methods within applied sciences and water planning in the North. NHF membership is open to all individuals and institutions/companies actively interested in hydrological work or research.

The purpose of the conference is to share experience in different fields of hydrological research and practice, and improving management of water resources. The conference will last three days and revolve around several themes with poster and oral presentations followed by discussions. Themes:

- Advanced Methods and Technologies in Hydrological Practice
- Agriculture, Forestry and water, Land-use changes
- Arctic Hydrology and Glaciers
- Eco-Hydrology in regulated rivers and streams
- Climate and Energy Systems (CES special session)
- Hydropower and Hydrology
- Uncertainty and Extremes in Hydrology
- Water Quality
- Water Resources Management under Climatic Change

The Conference is being organized by the National Power Company (NPC) and the Hydrological Service (HS) on behalf of NHF in co-operation with the Icelandic Hydrological Committee (IHC) and the University of Iceland (UI). The Organizing Committee is chaired by Óli Grétar Blöndal Sveinsson (NPC) and composed of Árni Snorrason and Jórunn Harðardóttir (HS), and Kristinn Einarsson (IHC). The Scientific Advisory Committee is co-chaired by Sigurður Magnús Garðarsson and Hrund Ólöf Andradóttir (Iceland) and composed of Jens Christian Reefsgard (Denmark), Bjørn Kløve (Finland), Agrita Briede and Elga Apsite (Latvia), Arvydas Povilaitis (Lithuania), Ingjerd Haddeland (Norway), and Gia Destouni (Sweden).

Any opinions, conclusions and recommendations expressed in these proceedings are those of the authors and do not necessarily reflect the views of the members of the Scientific Advisory Committee or the editors.

The organizers wish to thank those who helped make the conference possible and all the authors for their contribution.

PLENARY SESSION 1

NORTHERN HYDROLOGISTS AND THEIR GLOBAL ROLES

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ABSTRACT

Northern hydrologists play important roles in global hydrology-related research, education, activities in international bodies and programmes, and development co-operation. Attempts to quantify a "Nordic" level of activity indicate between 5 and 15 % of global totals. The paper's ambition is to illustrate and exemplify.

Regarding "impact sectors" there is an obvious emphasis on Northern specialties: Glaciers, snow and ice, lakes and hydropower, but even non-regional topics like stochastic hydrology, groundwater, droughts and floods, air pollution impacts, urban runoff and wastewater treatment have attracted much interest among Nordic hydrologists.

An interesting observation, probably universal, is the attraction exerted by water to make people of various professional backgrounds devote their energy to hydrology. Excellent contributions have thus been made by Nordic physicists, chemists, climatologists, geologists, botanists, - even physicians and diplomats -, all fascinated by water.

INTRODUCTION

Northern hydrology, however defined[1], will be expressed by *individual hydrologists*; in their research and their teaching, in writing and speech, in their interaction with others through working groups, symposia, and international development cooperation, or otherwise. It might therefore be relevant for the topic of the conference to illustrate how *individuals* have contributed to promote hydrology globally. The number of professionally active hydrologists today in the five strictly Nordic countries may probably be several hundreds, to which could be added former generations of colleagues. The number of their interactions makes it obviously impossible to

[1] In this paper the term *Nordic* (comprising Denmark, Finland, Iceland, Norway and Sweden) is used synonymously with *Northern*.

be complete and exhaustive in recording scientific contributions, institutions, projects and programmes, important symposia and other meetings etc.
Hence, the paper's ambition is to illustrate and exemplify.

SCIENTIFIC ACHIEVEMENTS

A straightforward way of measuring the possible *global role* of Northern hydrologists is by means of citation indices, which may, as the word implies, *indicate* the importance of a scientific paper. It is a fact, however, that hydrological research is not as much cited as for instance some medical publications. Probably the most-cited single article by Nordic hydrologists (either as main author or co-author) through the latest 7 years is *Hisdal* et al. (2001). The count is based on the ISI Web of Knowledge webpage, which covers nearly 15 000 journals, incl. Nordic Hydrology, http://portal.isiknowledge.com/portal.cgi . Other much-cited authors (>30 ISI items) include *Dan Rosbjerg, Poul Harremoës, Malin Falkenmark* and *Stein Beldring*. Judging from the topics of those much-cited works and authors, *(e.g.Hisdal et al. 2001, Brun et al. 2002, Beldring 2002, Morgenrot et al. 2002*), the impact seems to be considerable in such fields as for instance drought hydrology, wastewater treatment, and precipitation-runoff modelling.

A rather inaccurate, but interesting index, is a count of Nordic authors or co-authors among the references of major textbooks. One such example is *Arnell (2002): Hydrology and global environmental change*, containing a comprehensive list of 754 references, of which 37 are authored or co-authored by Nordic scientists. Is it fair to indicate that the Nordic contribution to global hydrology may be of the 5% order of magnitude? Cf. even similar counts regarding use of satellite imagery in snow and glacier studies (below)

A few hydrologists and their specialties are mentioned on the following pages:

Hydrometry. Examples of early Nordic inputs to global hydrology are noted within hydrometry. Maybe not surprising, as data collection is a natural early step in the development of a science, and long-term series are needed for water resource development. In most Nordic countries hydropower development was a driving force in developing hydrometric networks. One example of early contributions to hydrometry is the use of the instantaneous salt-dilution method for measuring discharge in turbulent streams. After developing the method in the mid-1920s, *Johan Aastad* and *Reinhard Søgnen*, who both were directors of the Norwegian hydrological service between 1921 and 1959, published their findings internationally *(Aastad and Søgnen 1954)*. Later, this method has been improved through automation and use of portable PCs, and it is much used in turbulent mountainous rivers in North America and Europe.

In more recent years, Nordic hydrologists have been in the forefront in developing and applying electromagnetic field measurements, whether in radar investigations of aquifer characteristics (*Niels Bøie Christensen),* and for measuring depth and water equivalent of snow *(Knut Sand and Oddbjørn Bruland 1998),* or using gamma-ray radiation or satellite imagery for snow cover and glacier mapping *(Martti Hallikainen, Risto Kuittinen 1989, Rune Engeset 2000).* The review article by *König et al (2001)* on measuring snow and glacier ice properties from satellites contains more than 160 references mostly from the 1990s. Among these, 14 percent have Nordic authors or co-authors. A slightly earlier review paper *(NHP Report 41 1996)* on the use of remote sensing for snow cover and precipitation estimates contains 55 references from Finland, Norway and Sweden.

There has always been a strong belief in our countries in the value of long-term hydrometric networks, reliable observations and sound data management. Ten years ago the number of hydrometric stations in the five Nordic countries was close to 4000, including more than 2000 discharge stations, *(Puupponen 1995),* most of which are today equipped with electronic loggers or telemetry. This average network density is remarkably high whether compared to the area of the region: 1 station per 310 km^2, or to the population of 24 mill.: 1 station per 6000 inhab. Puupponen has also coordinated network activities within the European region of WMO, aiming to improve the contribution of national hydrological services to implementation of the EU Water framework directive.

Snow and glaciers. Northern countries enjoy snowy winters and many glaciers. No wonder that Nordic scientists took up studies of snow and ice very early. Already in 1792-94, the Icelandic physician *Sveinn Pálsson* visited Iceland's glaciers ("ice mountains") and wrote in 1815 a scientific manuscript called Jöklaritiđ. In 1807, the botanist *Göran Wahlenberg* described glaciers in the Swedish mountains in a scientific way, and in 1824 the geologist *Jens Esmark*, based on observations of end moraines in Norway, speculated that Scandinavia had once been covered by glaciers. A few years later, 1837, the Swiss-American scientist Louis Agassiz, made his hypothesis of a global ice age[2]. The Nordic glaciological science as well as quaternary geology, developed quickly in the 1940-1960s by, among others, *Hans Wilson Ahlmann* and *Sigurđur Þorarinsson.* Ahlmann, who was Sweden's ambassador to Norway 1950-55, contributed to the classification of glaciers, based on field work also in Greenland. *Þorarinsson*, who was a leading geoscientist in the Nordic region, was even a prominent volcanologue, and studied jökulhlaups in Iceland. In this period the Iceland Glaciological Society (Jöklarannsóknafélag Íslands) was founded (1950). The

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[2] Agassiz's reputation as father of the Ice age hypothesis has later been contested, see e.g. http://en.wikipedia.org/wiki/Jean_Louis_Rodolphe_Agassiz

society issues the journal Jökull, still a leading international scientific journal in glaciology, and operates since 1951 a research station on Europe's second largest glacier, Vatnajökull. In Sweden, the mass balance studies of Storglaciären glacier started in 1946, constituting the longest continuous annual mass balance records in the world. The studies of Storglaciären have been facilitated by the proximity to the research station at Tarfala, established by *Valter Schytt*, and operated by Stockholm University. The mass balance records at the Norwegian glacier Storbreen comes second in the world, being started in 1949 by *Olav Liestøl (Liestøl 2000).* This glacier's front position has been continuously recorded since 1902.

Nordic glaciologists did not interpret the word "Nordic" narrowly, and organised 1949-1951 a Norwegian-Swedish-British Antarctic expedition to Queen Maud Land, inspired by *Ahlmann*. Those and later studies of the Antarctic, confirm that there has been no appreciable thinning of this part of the Antarctic ice, having strongly negative ice temperatures, in contrast to the recent retreat of most other glaciers of the world.

Nordic glaciologists have contributed much to promote glaciological field work and glacier surveys. The "Nordic" school of glaciology has based investigations on extensive field measurements of winter accumulation and summer ablation on a large number of glaciers. An early glacier inventory, which became a model for later publications, concerned glaciers in Northern Scandinavia (*Østrem et al. 1973*). Later examples are descriptions of European glaciers (e.g. *Schytt 1993*) and handbooks for field work (*Østrem and Brugman 1991*). *Gunnar Østrem* divided his professional career between Norway, Sweden and Canada. His main scientific finding is the realization that ice-cored moraines mainly consist of superimposed ice and not glacier ice.

The history of glaciological science is full of scientists coming from other fields of interest and making major contributions. One such is the Danish paleoclimatologist *Willi Dansgaard*, who in the late 1960s first demonstrated, using mass spectrometry, that the relative concentrations of hydrogen isotopes (H^1, deuterium and tritium) and oxygen isotopes, notably O^{16} and O^{18}, in ice cores and in trapped air bubbles can indicate climate changes, i.a. in the Camp Century, Greenland, ice core. The relative abundance of water molecules with different combinations of these isotopes is telling evidence of the temperature and humidity of the original air masses. Dansgaard's scientific achievements and efforts in the field work are told most entertainingly in his autobiography (*Dansgaard 2005*)

<u>Ice on lakes and rivers.</u> The discipline of ice on lakes and rivers has attracted many Nordic scientists. One pioneer in Nordic ice science was *Olaf Devik*, (*1932*), who was professionally active in this field until the 1980s when he was close to 100 years old. *Erkki Palosuo*, professor of geophysics in Helsinki and associated with the Finnish Institute of Marine Research, was

a pioneer in Finland's ice investigations, both in inland waters and the Baltic Sea. Other prominent names are *Arne Moberg*, *Juha Kajander* and *Edvigs Kanavins*, born in Latvia. Long time series of ice formation and break-up dates in rivers and lakes have acquired much interest in recent research on climate change.

Hydrological modelling. Within this field, Nordic hydrologists have contributed both in developing stochastic hydrology, using probability theory, and deterministic modelling. Stochastic hydrology was developed in the 1970s, quite early, internationally speaking, by among others *Eggert Hansen* who belonged to the staff of the Hydraulic Laboratory at the Technical University of Denmark, doing research on sediment transport in streams. The same group of scientists included also *Frank Engelund* and *Dan Rosbjerg*, all of whom contributed substantially to these studies. Among later contributors, *Lars Gottschalk*, Oslo University, should be mentioned. He received 2007 the Henri Darcy medal, which is awarded by the EGS / EGU (European Geosciences Union) *"in recognition of outstanding scientific contributions in water resources research and water resources engineering and management"*. Two of the three Nordic hydrologists who have received the medal: Falkenmark 1999, Rosbjerg 2001 and Gottschalk 2007, belong to the "Nordic school of stochastic hydrology", Cf. also *Lena Tallaksen et al.* (*1997*), who applied stochastic methods to drought studies, and was awarded the 1998 Tison Award. Among other Nordic groups active in hydrological modelling could be mentioned the Finnish Environment Institute (SYKE) where *Bertel Vehviläinen* is a leading scientist. There are also many examples of Nordic hydrologists forming international author teams on topics outside the Nordic region, e.g. *Haddeland et al. 2006*.

Erik Eriksson was the first professor of hydrology at Uppsala University, and has contributed much to the development of Swedish hydrology. His background interests were mainly within the chemistry of groundwater as well as meteorology and atmospheric chemistry. Together with Carl Gustav Rossby he helped to expand a regional network for atmospheric chemistry into the continent-wide *European Air Chemistry Network* in the mid 1950s, thus generating new insight into the long-range transport of air pollutants, (see below). He inspired younger hydrologists to specialise in theoretical hydrology, and use of mathematical models. One good example is the HBV model first tested in 1972, and developed 1976 by *Sten Bergström* in his PhD thesis, (*Bergström 1992*). The model is a distributed rainfall-runoff model, which includes conceptual numerical descriptions of hydrological processes at the catchment scale. It is a standard tool for Nordic hydrologists, and in different versions the HBV model has been applied in more than 50 countries all over the world, under different climatic conditions as for example Sweden, Zimbabwe, India and Colombia. The model has been applied to scales ranging from lysimeter plots to the entire Baltic Sea drainage basin, and for nationwide mapping as well as for modelling nutrient transport and impacts of climate change.

Acid precipitation. Water problems are manifold, and occur in all scales. Studies of the semi-global ecological problems in waters exposed to deposition of polluted air became a Nordic "specialty" in the 1970s, when damage to fish stocks and other aquatic life observed since about 1900 was linked to long-range transport of air pollution, LRTAP. The blame was put in particular on sulphur and nitrogen compounds, and "acid rain" became a catchword for this link. Nordic hydrologists took an active part in developing knowledge and counter-measures. The European Air Chemistry Network mentioned above was a basis for an OECD program, which during the 1960-70s established LRTAP as a scientific fact. The first UN global topical conference was organised in Stockholm 1972 on environmental problems, and ecological acidification was definitely put on the global agenda. An eloquent spokesman on these problems was soil scientist *Svante Odén,* another Uppsala professor, who played a role in waking up North American environmental scientists and politicians to the threat. A major step in scientific understanding, with strong inputs of hydrological catchment modelling, came with the Norwegian research program "Acid precipitation – effects on forests and fish " (*Lars Overrein et al*., *1980*). The political follow-up within the UN Economic Commission for Europe, ECE, was founded on the 1979 Geneva Convention on LRTAP, and has led to a series of protocols for reducing harmful emissions, the first one, on sulphur emissions, signed 1985 in Helsinki. Later protocols to the LRTAP convention have been signed in Oslo (1994), Aarhus (1998) and Gothenburg (1999). Even in this international political / legal setting, Nordic hydrologists have taken part, notably *Arne Tollan* and *Lars Nordberg*, both serving as heads of the ECE secretariat unit responsible for negotiations related to air pollution, (*Sliggers and Kakebeeke 2004*).

Hydrological knowledge is an undisputable foundation for good management of water resources. Without that basis, the global water crisis, or rather crises can hardly be solved. One Nordic hydrologist in particular, *Malin Falkenmark*, has built bridges between the water sciences and the political world. She has described the linkages between land use - other human impacts – water characteristics – and environment. One useful concept introduced by Falkenmark is the terms *"green and blue water"* to better assess the role of water in plant production. *Green* water is defined as the fraction of water that is evapotranspirated, i.e. the water supply for all non-irrigated vegetation. Green water is either productive with respect to plant production (if transpired by crops or natural vegetation) or non-productive (if evaporated from soil and open water). *Blue* water refers to the water flows in groundwater and surface water (river, lakes). It represents the water that can be withdrawn e.g. for irrigation or is available for *in-situ* water use. A brief presentation is available in *Falkenmark and Rockström, 2005*.

Falkenmark, who has been associated with SMHI, various Swedish universities, the Natural Science Research Council, and SIWI has received several prestigious prizes for her work: The KTH price 1995; the International hydrology prize 1998; The Volvo environmental prize 1998 (with D. Schindler); and the Henry Darcy medal (EGU) 1999. In 2005 she received the Crystal drop award of the International Water Resources Association, as well as the Rachel Carson prize. Quoting from one presentation of Malin Falkenmark: "She is one of a small group of analysts of large-scale global and regional water problems whose work with broad perspectives has been instrumental in raising the profile of water issues internationally." [3]

Although Malin Falkenmark has a special position among Nordic hydrologists studying the looming global water scarcity, there are also others who contribute in the international discussion on sustainable water management in developing countries, for instance *Olli Varis* of the Helsinki University of Technology, whose background is in water quality modelling. See e.g. *Vakkilainen and Varis 1999*. Another person to mention is *Torkil Jønch-Clausen*, director at the Danish Hydraulic Institute. He has held several central positions (cf. annex 2), and was the 2006 winner of the Hassan II Great World Water Prize.

Urban hydrology research in the Nordic region has been quite active in the densely populated parts of southern Scandinavia, e.g. the Technical University of Denmark (Copenhagen), Chalmers Technical University (Gothenburg), and Lund Technical University. There are also active groups concentrating on "cold cities" at the Norwegian University of Science and Technology (Trondheim), and the Luleå University of Technology. Some important exponents for urban hydrology research at these centres of water technology are *Poul Harremoës, Gunnar Lindh, Jan Niemczynowicz, Lars Bengtsson, Sveinn Thorolfsson,* and *Angela Lundberg.* Modelling of snowmelt in cities is obviously of particular interest in our region, and is receiving much attention. On the side of practical urban design, Northern researchers have been strong proponents for local infiltration of surface water as a best management practice, BMP.

Numerous Northern scientists have contributed much to other sectors of hydrology than those mentioned above, not least in describing national or regional water conditions, and management applications. The Nordic countries are energy-rich, and hence much research has been carried out on the interface of climate change and renewable energy, notably hydropower. Some of this research has been organised as joint Nordic programmes and

[3] http://www.worldfoodprize.org/symposium/2002/2002spkrbios.html#falkenmark

supported by the Nordic Council of Ministers or the Nordic national hydrological services (e.g. *Nils Roar Sælthun* et al., *1998* and *Jes Fenger 2007*. See also *Risto Lemmelä and Nea Helenius 1998* and *Arni Snorrason et al. 2002*).

Other fields of strong Nordic professional interest are e.g. groundwater and other sub-surface hydrology, flood hydrology, lakes and wetlands, and fluvial geomorphology, *(*e.g. *Sundborg, 1957)* all of which would have deserved closer analysis.

SYMPOSIA AND CONFERENCES

Contributions to the scientific progress often materialize as papers presented at symposia and conferences, but even in the very organization of international professional meetings. For obvious reasons the number of domestic participants at symposia will be high, and thus provide ample opportunities for presenting research results from the actual region. This section deals with international hydrological conferences and symposia, held in the five Nordic countries for an international audience. The responsible supporting organisations are mostly IAHS, Unesco or WMO, occasionally all three together, often guiding the choice of scientific topics, and taking care of the publication of proceedings. Every year several water-related international meetings of interest for professional hydrologists are organised in the Nordic countries. Unfortunately, it would be practically impossible to record all.

Instead, a survey of *some* of the many important meetings of this kind since ca. 1950 is given in Annex 1. Not surprisingly, the topical emphasis is on what many will identify as Northern specialties, like snow, ice and glaciers, lake hydrology and mountainous regions. Successful meetings often foster repetitions, which may grow into a regular series of symposia / conferences. The annual World Water Week in Stockholm, organised by the Stockholm International Water Institute, SIWI, is one example, with inputs from several hydrologists, like *Ulf Ehlin* and *Malin Falkenmark*. The series of International conferences on Climate and water, held with 9 years intervals in Helsinki 1989, 1998, and 2007, is another. In this respect *Esko Kuusisto* has provided leadership.

The International Association for Hydraulic Research, IAHR, deals with ice science and engineering, among other fields. Since 1970, the leading forum for ice issues is the biannual IAHR International symposia on ice. Over this period the symposia have been held in the Nordic countries a number of times: Reykjavik 1970, Luleå 1978, Espoo 1990 and in Trondheim 1994 (IAHR 1994, and similar previous proceedings).

PUBLICATION AND EDUCATION

Nordic Hydrology The scientific journal Nordic Hydrology may illustrate the ambition of Northern hydrologists to play a global role. The journal was initiated through Nordic IHD cooperation in the 1960s as a publishing ground for hydrological research "*in its widest sense",* and was intended to act as a window for Nordic hydrologists to the worldwide hydrological community. The journal has existed for almost 40 years, since 1976 owned by the Nordic Association for Hydrology, Financial support from Nordic research councils ceased after 2000, at the same time as the need for electronic publishing became urgent. A joint ownership with IWA Publishing was established in 2004, when also a web based version of the journal was launched. From 2008 the British Hydrological Society will enter into the cooperation, and the journal will change name to Hydrology Research. The journal is covered by the Science Citation Index and the contents are widely abstracted. All papers are internationally peer reviewed. Editor today is *Dan Rosbjerg*, who succeeded *Arne Forsman* and *Eggert Hansen* in that position. From 2008 Ian Littlewood, UK, will become co-editor.

Textbooks. Whereas there are plenty of water-related textbooks in Nordic languages, there are, understandably, few written in other languages. Some examples of Nordic-authored books which have made it to the student's library: Falkenmark initiated and co-edited the book "Comparative Hydrology" published by UNESCO (*Falkenmark and Chapman 1989*). The book on hydrological drought by *Tallaksen and van Lanen (2005),* is also being used as a textbook, as well as a book series on Hydropower development, *NTNU (1992-2003). Østen Tilrem*'s five-volumes Manual on procedures in operational hydrology, *(Tilrem 1997)* is based on practical experience in development projects in Africa, and is still used in training of hydrological field personnel and technicians from developing countries, both in Norway and Sweden (SMHI courses).

International programmes and training courses are available at a large number of Nordic universities, and attract large numbers of students from all over the world. Distance education tools are often applied. Some examples are mentioned here.

"The United Nations University Geothermal Training Programme" is operated by *Orkustofnun* (the National Energy Authority of Iceland) under a special agreement with the United Nations University.

"Water Governance in Long-Term Perspectives", supported by NordForsk (Nordic Research Board), is a Nordic-Baltic interdisciplinary research training course, given at the *University of Tampere*.

"International Centre for Hydropower", ICH, Trondheim, supported by Norad, The Norwegian Agency for Development Cooperation, offers international training courses and other information activities promoting development and use of hydropower resources.

The "*Norwegian University of Science and Technology"* (NTNU), also in Trondheim, offers a two-year international MSc programme in hydropower planning. A unique tool for this kind of education is a 17-volume textbook series on hydropower development, issued over the time period from 1992 to 2003. There are separate volumes on Hydrology and on Environmental effects, *(NTNU, 2003).*

There is a trend to create interdisciplinary groups of several universities of various countries, offering water-related education. Some examples:
"Universities' partnership for Transboundary waters" Linköping university (Jan Lundquist) is one of nine partners in this consortium.
Another example is the Erasmus Mundus European "*Joint Master in Water and Coastal Management",* since 2004, where the *University of Bergen* participates together with universities in Portugal, Spain and the UK.
A third example is "*The International Research School of Water Resources"* , a formal collaboration between nine Danish universities and research institutes with expertise within water resources. Director is *Karsten H. Jensen*, Professor at the University of Copenhagen. The *"Baltic University Programme",* offering courses in sustainable water management has been implemented since 1998 at 30 universities in the Baltic Sea region. (*Lars-Christer Lundin 2004*)
International training courses are being held at various times and by various Nordic organisers in e.g. glacial hydrology, irrigation in arid zones, hydrometry, and more.

INTERNATIONAL ORGANISATIONS AND PROGRAMMES

Many Nordic hydrologists have over the years held positions in international water-related bodies, both as employees and in honorary posts. Individuals of course bring along into their work their professional competence and values, their opinions and working habits, their negotiating skills, as well as general human qualities. To the extent that such characteristics are coloured by one's region of origin, it will be fair to speak about a global role for Northern hydrologists. An indication of such influence is the fact that today 6 out of 63 elected officers of the IAHS and its 9 commissions are Nordic hydrologists, i.e. close to 10 %.

The following comments and Annex 2 are by no means exhaustive, but may provide valid examples of such global roles. In addition, a large number of Nordic hydrologists have served individually or as members of various committees and working groups of UNESCO, WMO, IAHR, IASC (Intl. Arctic Sci. Comm.), FAO, UNEP, World Bank, ICOLD, IUCN, IWHA, World Water Council, Regional developm. banks, IWA, and certainly many others.
The author appreciates any improvement and additions to the Annex 2.

The International Hydrological Decade (IHD) programme 1965-1974, launched by Unesco in cooperation with other UN agencies, was a major step for international hydrology, and was succeeded by the International Hydrology Programme (IHP). In each of the Nordic countries it meant a consolidation of hydrological investigations and networking. Between the Nordic countries, the IHD and IHP have led to a wide range of cooperative efforts in:

- *information exchange*: among others the international journal Nordic Hydrology (see above); the inter-Nordic hydrological bulletin "Vannet i Norden", founding of the Nordic Hydrological Association in 1970; publication of a Nordic Glossary of Hydrology *(Johansson 1983),* linked through English to the Unesco-WMO International glossary of hydrology.
- *joint research efforts,* within e.g. groundwater and soil moisture hydrology, water chemistry, snow measurements, water balance in representative basins, water balance of the Baltic Sea, and water data management. The FRIEND project (Flow regimes from international experimental and network data sets) has been a part of IHP since 1985. The project, which now has global coverage, started as a practical cooperation between four NW European countries: UK (Institute of Hydrology), the Fed. Rep. of Germany, the Netherlands and Norway, later joined by i.a. Sweden and Finland. It is a tool for collecting and analysing data from all participating countries. A central person during the first decades of extending the FRIEND project was *Lars Roald (Roald et al. 1989),* but several Nordic colleagues have spent time working in the core team, and numerous research papers have benefited from the FRIEND data base.

  Recently, scientific interest in the polar regions and their role as agents and indicators of climate change has materialized in the International Polar Year, IPY 2007-2008, organized by ICSU and WMO. Consolidation of the hydrology-related project proposals for the IPY ha has been effected by *Arni Snorrason*, into an Arctic Hydra programme sector.
- *joint Nordic representation* and development of joint positions in international forums such as the Intergovernmental Council of IHD and IHP, where a rotational seat is agreed, i.e. Sweden 1975-76, 1988-91, 2002-05; Finland 1977-78, 1992-95, 2008-11; Denmark 1979-83, 1996-99; Norway 1984-87, 1998-2001; Iceland 2004-07. It is a general opinion that such coordination has provided Northern hydrologists a stronger voice in international hydrology matters, *(Tollan 1974).*

DEVELOPMENT COOPERATION

Nordic hydrologists today take an active part in water-related development cooperation:

- as individual experts serving on projects financed and managed by national aid agencies;

- as individual experts recruited by NGOs, consultancy firms or international organisations;
- through institutional cooperation;

How did hydrological development cooperation start? There are obvious links between decolonization of the world after 1945 and development cooperation. The UN and its many organizations and programmes soon took up technical assistance projects. Concepts like *developing country, least developed countries, North-South axis, humanitarian aid*, *the 3rd world,* represent a perspective of the rich, industrialized Western world supplying knowledge and technology to the poor, rural peoples of the world. Other driving forces on the same arena were the moral call behind missionary work, and emergency relief assistance, all requiring water-related expertise. Anyway, development projects attracted young hydrologists ready to give technical assistance, designing station networks, organising fieldwork, managing water data, providing design data for dams or irrigation structures and teaching hydrology.

Jakob Otnes was a pioneer in this field of work, serving for FAO in the Rufiji basin, Tanganyika 1955-1960, seven years before the Norwegian agency for development aid was established, named Norad since 1968. One emphasis, then and later, in the Nordic hydrological support to developing countries has been development of hydrometric station networks and field work procedures. The immediate purpose has often been to provide design data for use by the hydropower sector, (e.g. *Tilrem 1997*). Otnes inspired others to follow in his track, and over the next 50 years, some 40 Norwegian hydrologists have served for shorter or longer periods in development projects, predominantly in African countries. Assuming proportionality, this could mean that some 150 Nordic hydrologists have made similar contributions. Prominent names from younger generations include e.g. *Torkel Jønch-Clausen* (D), *Jan Lundquist* (S), *Kjell Repp* (N), and *Joakim Harlin* (S).

Since the early 1990s there has been a shift in international thinking concerning water management from supply-based to demand-based management, being reflected also in international development cooperation. Key words in this new paradigm are *river basin management*, *conflict resolution, water legislation* and *integrated water resources management*.

Hydrologists now regularly take part in broad team work involving lawyers, economists, other social scientists, and engineers. In recent years, and particularly since formulation of the Millennium development goals in 2000 and their further elaboration at the WSSD in Johannesburg 2002, the emphasis has grown strong on water supply and sanitation projects, meaning new roles for the hydrologist abroad. It seems that success or failure of water-related projects often depend on factors outside the scientific-technological field, like internal or external politics, poor communication, and short-sighted commitment. Success stories often involve strong personal commitment, idealism and enthusiasm, *(Tollan and Repp, 2002).*

Some Northern countries have established national networks for mobilising water-related competence and disseminating knowledge internationally. Examples are the *Swedish Waterhouse* (http://www.swedishwaterhouse.se/) and the *Danish Water Forum* http://www.danishwaterforum.dk/ .

Although national statistics may be based on varying criteria, some comparison illustrates the importance of Nordic water-related aid. It has not been possible to assess the total support to the water sector from Finland and Iceland.

*Denmark**:*** Danida**,** Danish International Development Assistance. In 2006 Danida spent 8,6 % (=672 million DKK) of its bilateral aid on water supply and sanitation projects. http://www.um.dk/da/menu/Udviklingspolitik/

*Finland**:*** Ministry for foreign affairs, Department for development policy http://formin.finland.fi/public/default.aspx?nodeid=15316&contentlan=2&culture=en-US

*Iceland**:*** ICEIDA, The Icelandic International Development Agency, http://www.iceida.is/english. ICEIDA, in cooperation with Malawian authorities, has initiated a water and sanitation project, which will run until the end of 2010. ICEIDA's contribution will be 2.6 million dollars. Iceland's water-related development aid is moreover quite focussed on the fisheries and the geothermal sector.

*Norway**:*** Norad**,** The Norwegian Agency for Development Cooperation, http://www.norad.no/. In total, aid within water supply and sanitation make up a small proportion of the Norwegian bilateral development cooperation. The average during the period 1999-2003 was nearly 2,4 percent, equivalent to NOK 196 million per year, of a total bilateral development cooperation of NOK 8 240 million per year in average. The proportion of the water aid to Africa has been reduced to one fourth during the same period, *(ForUM, 2005)*. Norad provides considerable support to the energy sector, including hydropower.

*Sweden**:*** SIDA, Swedish international development cooperation agency, http://www.sida.se/, in 2006 spent 1,25 billion SEK on so-called sector programme support, and 22 % of this (=275 million SEK) went to the natural resources sector, where water-related projects are parts. 88 million SEK went to the water supply and sanitation sector in Kenya and Uganda.

CONCLUSIONS

Northern hydrologists have played, and continue to play, important roles in global hydrology-related matters. Quantifying such impact through quotation indices etc. is uncertain, but adds support to the statement above. Some such attempts at quantification indicate a “Nordic” level of activity between 5 and 15 % of global totals. The impact is many-sided, including research, education, activities in international bodies, programmes and forums, and development cooperation, with emphasis on operational and applied hydrology.

Hydrological knowledge is an indispensable basis for water resources planning and management. Interdisciplinarity is therefore increasingly important, as reflected particularly in the sectors of water education and water-related development cooperation. The need for sustainable water management in order to meet global water problems has stimulated important Nordic interest.

Regarding “impact sectors”, there is for obvious reasons an emphasis on Northern “specialties”: glaciers, snow and ice, lakes, and hydropower. However, even non-specific topics like stochastic hydrology, groundwater hydrology, droughts and floods, air pollution impacts, urban runoff and wastewater treatment, have attracted much attention among Northern hydrologists, and with good results.

An interesting observation, which probably is universal, is the power of attraction exerted by water, to make people of highly different backgrounds devote their professional energy to hydrology. Excellent contributions have been made by Nordic physicists, chemists, climatologists, geologists, botanists and physicians, - even diplomats. We are all fascinated by water.

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Several colleagues have given valuable comments to this paper, based on their own knowledge of Nordic hydrologists and their international contributions. I am pleased to recognize with much appreciation numerous suggestions from *Randi Pytte Asvall, Klas Cederwall, Lars Gottschalk, Hege Hisdal, Lars Nordberg, Markku Puupponen, Dan Rosbjerg and Rune Rosland.* The list of references includes several examples of joint Nordic-nonNordic authorship of major publications; another indication of global professional impact.

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Annex 1

Non-exhaustive list of selected major international hydrology-related symposia and conferences organised in the Nordic countries, ca. 1950-2007:

Year	**Symp./Conference**	**Place**	**Organiser**	**References**
1948	IASH General Conference	Oslo	IASH	AIHS. 1948
1966	Ground water problems	Stockholm	Swedish IHD comm.	Eriksson et al. 1968
1972	Distribution of precipitation in mountainous areas	Geilo	WMO/IAHS	WMO, 1973
1973	Hydrology of lakes	Helsinki	IAHS/Unesco/WMO	IAHS, 1973
1984	Third Intl. Conf. on Urban Storm Drainage	Gothenburg	IAHR / IAWQ	
1989	FRIENDS in hydrology	Bolkesjø	IAHS	Roald et al. 1989
1989,1998, 2007	Intl. conferences on climate and water	Helsinki	SYKE and others	Lemmelä and Helenius, 1998
1990	Arctic hydrology	Longyearbyen	Norw. IHD comm.	Gjessing et al. 1991
1991	Ecosystems approach to water management	Oslo	UNECE	UNECE, 1991 Tollan, 1992
Annually since 1991	World water week	Stockholm	SIWI	SIWI: Annual syntheses
1992	Erosion and sediment transport monitoring	Oslo	IAHS	Bogen et al. 1992
1994	Ice	Trondheim	IAHR	IAHR, 1994
2000	Extremes of extremes	Reykjavik	IAHS	Snorrason et al. 2002
2001	The role of water in history and development	Bergen	Intl. Water History Association, IWHA	Tvedt et al. (2006)
2005	Hydrology, ecology and water resources in headwaters	Bergen	IAHC, IAHS, Unesco, and others	NVE, 2005
2007	Pasts and futures of water	Tampere	IWHA	http://www. envhist.org/ bookofabstracts. pdf

Annex 2

Non-exhaustive list of positions held by Nordic hydrologists in selected international hydrology-related organisations, programmes and bodies ca. 1950-2007 (excl. development aid):

Name	**Organization**	**Position**	**Period**
Arheimer, B. *(S)*	IAHS, Comm. Water Quality	Vice President	2007-09
Boegh, Eva *(D)*	IAHS, Coupl. Land-Atmosp. Syst.	Secret. / Vice president	2003-07
Bogen, Jim (N)	IAHS; Commission on Continental Erosion	President	2005-09
Ehlin, Ulf (S)	Helsinki Commission Stockh. Intl. Water Institute, SIWI	Exec. secretary Dir./ Scient. Dir.	1994-96 1997-07
Forsman, Arne (S)	WMO, Hydr. and Wat. Res. Dept.	Chief	1965-67
Harlin, J. (S)	UNDP	Wat.res.spec.	2004-
Hisdal, Hege (N)	IAHS, Comm on Surface Water	Secret. / Vice president	2003-09
Hock, Regine (S)	IAHS, Snow and Ice	Secretary	2005-09
Jønch-Clausen, Torkil (D)	Global Water Partnership, Techn. Comm. Intl. Water Resources Assoc.	Chair Secr. general	1996-03 2005-06
Kuylenstierna, Johan (S)	WMO, HWR Dept. SIWI – Swedish Water House UN Water, FAO	Prof. off., Project Director Chief Techn. Adv.	1999-03 2003-07 2007-
Lemmelä, R. (F)	UNESCO, Bureau of IHD Co-ord. Council	Member	1995-96
Melder, Ole M. (N)	WMO, HWR Dept.	Project off.,	1964-93
Mustonen, S. (F)	IAHS, Comm. surface water	President	1975-83
Nordberg, Lars (S)	UN Economic Comm. Europe, UNECE IUAPPA + UN Air Poll. Progr./C.Asia	Head, Air Poll. Unit Adviser	1992-00 2000-
Rosbjerg, Dan (D)	IAHS	Vice president	2007-09
Sandholt, Inge (D)	IAHS, Comm. Remote Sensing	Vice president	2003-07
Slettenmark, Gustaf (S)	International Association for Scientific Hydrology, IASH	President	1947-48
Snorrason, Arni (I)	WMO, Inter-commission task force on IPY, and subcom. of the IPY Joint Committee	Member	2005-
Tollan, Arne (N)	UN Economic Comm. for Europe, UNECE European Environment Agency, EEA	Head, Air Poll. Unit National expert	1983-85 1992-93
Wingård, Bo (N)	UNESCO, Bureau of IHD Co-ord. Council	Member	1984-85
Wold, Bjørn (N)	International Glaciological Society	President	1993-96

CLIMATE CHANGE AND VARIABILITY: HOW CAN WE INFORM STRATEGIES FOR ADAPTATION

Upmanu Lall

ABSTRACT

Anthropogenic climate change has emerged as a major environmental concern In the 21st century. There is clear evidence that global warming has occurred in the 20th century, and its connection to the use of fossil fuels and greenhouse gas emissions is now widely accepted. The Intergovernmental Panel on Climate Change (IPCC, http://www.ipcc.ch/) has developed many scenarios for greenhouse gas emissions corresponding to different degrees of mitigation of greenhouse gas emissions. These scenarios have been used with more than 45 simulations of different coupled ocean-atmosphere general circulation models (GCMs) to develop projections of climate up to the year 2100, for each of the scenarios. The different simulations are made to account for uncertainty in the knowledge of climate model parameters and also because, since the equations governing climate are nonlinear, small differences in the starting values can potentially translate into large differences in the resulting climate statistics. Climate change impact assessment and climate change adaptation work is then informed by the rainfall and temperature data from these simulations.

The talk will first review some hydrologic aspects of these simulations And highlight the challenge faced in directly using such "data" for water resource management and project design. The argument is made that substantial "processing" of these simulations is needed to correct biases and such, in the control runs covering the 20th century, suggesting that as far as hydrology is concerned the model physics is likely to be far from the real world physics. At one level this argues against the use or credibility of these models for the future. At another level, the question becomes whether there are any state variables in these models that could be used to inform key variables of interest in water management, even in a probabilistic sense.

This idea is developed through an example of what non-stationarity means for risk based or probabilistic analysis as typical in our field, using retrospective analyses constrained perhaps by long term proxy paleoclimate data that exhibits change over many time scales. The prospective application for the future is then outlined.

NORDIC HYDROLOGY IN CLIMATE CHANGE CONTEXT: THE OURANOS EXPERIENCE IN QUÉBEC

André Musy, Executive Director, Ouranos, Montreal
René Roy, Hydrologist, Hydro-Québec Montreal

ABSTRACT

OURANOS is a research and development consortium focusing on regional climatology and adaptation to the climate change. Created in 2001, following major disasters in Quebec due to exceptional climatic events, OURANOS' mission is to develop our knowledge of climate system, and to establishing plausible climatic scenarios using climate simulations carried out by different regional and global circulation models. These results allow undertaking impact and vulnerability studies related to several important social issues (such as coastal erosion, integrated water management, natural resources, population security, natural environment) and are essential for implementation of appropriate adaptation climate change strategies.

The northern Quebec is particularly affected by the effects of these potential changes; especially affected is the permafrost regions, the streamflow of northern rivers, duration and annual cycle of freezing/thawing. The consequences concern the use of natural resources, especially for hydroelectric power generation, as well as people's security and the maintenance of the fragile ecosystems located in these regions. New methodological approaches should be considered, especially in terms of water flow on partially frozen land whose behaviour is becoming particularly complex for organic soils.

Ouranos has launched, with some of its faculty (INRS-Québec, Laval University) and institutional members (Hydro-Quebec) several studies focusing on Nordic hydrology. The issues of water behaviour on peat soil in thawing, as well as the quantification of water inflow in northern Quebec’s dams are considered. Several specific projects are developed in these directions and are supplemented by others, like dendrochronological studies whose objective is to explore the past hydrology in order to better project its future.

The presentation will address these issues and present the methodological approaches regarding simulations and climate scenarios as well as the results achieved concerning the runoff occurrence in the northern regions of Quebec.

URBAN HYDROLOGY AND SUSTAINABLE URBAN DRAINAGE AT LAKE URRIDAVATN IN GARDABAER, ICELAND

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ABSTRACT

This paper describes the comprehensive study of the urban hydrology (UH) and the selection of stormwater management solutions at Lake Urriðavatn. Urban hydrology involves both water quantity and water quality. Lake Urriðavatn is located in the municipality of Garðabær, ca 15 km southwest of the capital of Iceland, Reykjavik. According to the Master plan for Garðabær, 1995–2015 the area is to be urbanized, while the lava peninsula north of Lake Urriðavatn is to be protected against urbanization.

Lake Urriðavatn is rich on vegetation, fish and bird life. The goal for Lake Urriðavatn is to keep it as untouched as possible, by taking care of the water and the land environment in the watershed. By selecting Sustainable Urban Drainage Systems (SUDS) the goals may be achieved and the natural water resources of Lake Urriðavatn will be acceptable, but the water quality Class B has to be accepted for some parameters.

The IKEA lot on the lava peninsula north of Lake Urriðavatn is finished implemented with SUDS solutions and it is functioning well Lake Urriðavatn.

Keywords: Integrated stormwater management (ISM), Lake Urriðavatn, Stormwater Management (SWM), Sustainable Urban Drainage Systems (SUDS), Urbanhydrology (UH), Urbanization.

INTRODCUTION

Lake Urriðavatn is located in the municipality of Garðabær, ca 15 km southwest the capital of Iceland, Reykjavik. According to the Master plan for Garðabær, 1995–2015 the area at Lake Urriðavatn is to be urbanized. The plans include new homes, businesses, retail parks, service centers and industrial areas. In addition the plan includes competence centers and schools, university and research buildings etc.

Lake Urriðavatn is rich on vegetation, fish and bird life, which is not to be degraded during the urbanization. Lake Urridavatn itself and its nearest surroundings are protected from all disturbances. The lava peninsula north of Lake Urriðavatn is also protected against urbanization, because of its biological diversity and recreation.

The goals for Lake Urridavatn are to keep it and its surroundings as untouched as possible. That means that the water quality class in Lake Urridavatn should unchanged and be as near Class A, untouched water, as possible. Therefore the ambitions on environmental friendly planning and implementation are high and the creations of high natural and high quality of living standards in the watershed are prioritized.

I n the planning process, the need for stormwater management resulting water quality and quantity control became apparent, and the developer, Þekkingarhúsið EHF, announced an international competence on planning and development of stormwater management systems that were appropriate for the watershed. This competence was won by COWI LtD, Norway in cooperation with the author. This project has been conducted in collaboration with several partners in Iceland, such as Alta LtD, Landslag LtD, VST LtD, Hnit LtD, Municipality of Garðabær, the local pollution authority, the local health authority, ISOR (Íslenskar orkurannsóknir) and Háskólasetur í Hveragerði, and international partners such as the UK Architects John Thompson & Partners, London, UK. Alta LtD has been the coordinator.

Stormwater management includes the principles of urban drainage, but it includes also wider environmental issues. They are of significance not only to engineers, but also to such as practitioners in environmental science, ecology, urbanhydrology, technology, policy and planning, geography, health studies etc.

Sustainable Urban Drainage Systems (SUDS) is a relatively new approach in urban drainage, where the stormwater is infiltrated and retained in the catchment and considered as a positive resource in the urban landscape, where aesthetics, multiple use and public acceptance of chosen technical solutions are important. SUDS are able to treat the stormwater and remove pollutants before the discharge of the stormwater, Stahre (2006) and Mays (2001).

Integrated stormwater management (ISM) is also a term to describe this new approach. The term integrated underlines the multidisciplinary interest and the active involvement by different technical departments in a city such as planning, park & recreation, environment, drainage etc. The characteristic feature of the new approach is that ecological processes are involved, and utilization of the nature's capability of handling water. Such processes are; flow equalization in open water courses, infiltration and percolation into the ground, fixing and biodegradation of pollutants in upper ground-level and uptake of pollutants by vegetation, Stahre (2006) and Mays (2001).

This paper describes the stormwater management solutions that were suggested by the Norwegian team. The Norwegian team prepared a report with the title: Stormwater management at Urridaholt, Retail Park – Best Management Practices. Final report 08 2005. Authors; Asle Aasen, Svein Åstebøl, Sveinn Thorolfsson, Aasen et al. (2005)

The solutions are based on the goals for water quality in the lake and a control of pollutants in the watershed, taking care of water and land environment. By use of Sustainable Urban Drainage Systems (SUDS), the goals for Lake Urriðavatn may be achieved.

THE STUDY AERA

Lake Urriðavatn is a small, shallow lake at 30 masl. The size of the watershed is 235 hectares i.e. 2.350.000m2, where 460.000m2 are protected area around Lake Urriðavatn, of which 200.000m^2 are wetlands.

Geography and urbanization

There are three hills in the watershed i.e. Urriðaholt in east, Setbergsholt in west and Hádegisholt in south compared to Lake Urriðavatn. They arise up to 65 and 125 masl.

After the urbanization the area distribution between urban and not urbanized areas in the watershed draining into Lake Urriðavatn from different catchments, are shown in table 1.

Table 1. Future area distribution in catchments after the urbanization (area in hectares, ha).

Catchment no.	U-1	U-3	IB-27	IB-28	Sum	%
Green areas in the urbanized areas	8.5	18.0	8.9	6.5	41.9	46%
	8.9				8.9	10%
Impervious areas in other developing areas		22.0	10.9	7.9	49.4	44%
Sum urbanized areas	17.4	40	19.8	14.4	91.6	39%
Not urbanized areas					149.1	61%
The whole watershed					235	100%

Legend: U-1 is the IKEA lot, U-3 is Urridaholt and IB-27 plus IB-28 are Setbergsholt (holt = hill)

The firm, IKEA applied for a building permission for a shopping mall at the lava peninsula north of Lake Urriðavatn. It initiated an evaluation of the future runoff conditions and the potential pollution in the whole watershed to Urriðavatn. The study resulted in the acceptance of the building permission under strict requirements on environmental friendly planning and sustainable stormwater management at the IKEA lot and also in other catchments.

Geology and hydrology

In the peninsula north of the Lake Urriðavatn, the geology is characterized by a thick lava layers with high porosity. In Urriðaholt in the east there is tight rocky.

Lake Urriðavatn was formed 7-8000 years ago, when the lava field, Búrfellshraun was flowing and blocked the outlet form the valley. Today there is a small earth dam at the outlet, controlling the outflow.

The surface of Lake Urriðavatn is 140.000m^2 (14 ha) and ground with low porosity, and with thin layer of sandy soil at the top and similar conditions are in Setbergsholt west of the lake. Between these hills and the lake are thick layers of boggy ground the average depth is only 0.5–0.6m. The volume is calculated to 77.000m^3. South of Lake Urriðavatn is a protected wetland, named Dýjamýri, discharging ca. 20l/s into Lake Urridavatn through a constructed creek, Dýjakrókalækur, Hjartarson (2005). No other creeks feed water into the lake, but springs feed Urriðavatn with natural discharge. Some of these springs originate from a long distance. in the peninsula north of the Lake Urriðavatn

The only visible runoff out of Urriðavatn is via a small creek called Stórakrókslækur, with an average flow of about 50l/s. In addition ca. 30l/s are calculated to run through the lava north of the lake, Hjartarson (2005).

Climate and urban hydrology

Despite the north Atlantic location, the Reykjavik region is not as cold as expected. The region is a very wet, having more than 200 rainy days every year. According to Køppen’s classification, Reykjavik region has maritime, coastal, subarctic climates, Køppen (1936).

Table 2 shows the monthly and annual values on precipitation and temperature in the City of Reykjavik and at Vifilstaðir. Vifilstaðir is located 4 km east of Urriðavatn, Jónsson (1986).

Table 2. Precipitation in Reykjavik and at Vifilstaðir 4 km east of Urriðavatn (mm/yr).

Location	Jan	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec	Year
Reykjavik	76	72	82	58	44	50	52	62	67	86	73	79	799
Vifilstaðir	95	114	113	66	54	53	55	83	83	128	101	90	1035

Table 2 (continued). Temperature in Reykjavik and at Vifilstaðir (°C).

Location	Jan	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec	Year
Reykjavik	-0.9	0.5	1.1	3.3	6.2	8.6	10.6	10.2	7.4	4.6	0.8	-0.5	4.3
Vifilstaðir	-1.2	0.3	1.0	3.5	6.5	8.9	11.0	10.3	7.4	4.5	0.5	-0.8	4.3

The table 2 shows a great difference in annual precipitation in Reykjavik and at Vifilstaðir or 336mm/yr. The temperature at Vifilstaðir is slightly higher in summer than in Reykjavik but in autumn slightly low. The developer, Þekkingarhúsið EHF, is keen on the winter runoff conditions in the area, Thorolfsson, (2004).

In urban hydrology time and area scale are different from that in hydrology in rural catchments, i.e. minutes or hours vs. days and months and m^2 and hectares vs. km^2. The urban runoff is characterized by a quick response compared to rural runoff, because of impervious surfaces, lined surfaces, vegetation removal and drainage facilities such as pipes etc for quick conveyance of the stormwater out of the catchment.

When planning and designing urban stormwater facilities data on short-term rainfall are required. Jónasson et al. (1994) has analyzed the precipitation data in the Reykjavik region regarding designing stormwater facilities and presented so-called M5-curves. These curves can be used to design stormwater facilities at Urriðavatn. Such IDF-curves are presented in Aasen et al. (2005) prepare by Sigurdsson (2004) for the Urriðavatn area, based on M5 = 53.5 mm/day.

STORMWATER MANAGEMENT AT LAKE URRIDAVATN

Goals and criteria for stormwater management are to relay on regulation and requirements to protect Lake Urriðavatn, Umhverfismálaráðuneyti Íslands (1999). The following goals and criteria were prepared for stormwater management in urbanized areas:

- The stormwater is to be managed locally i.e. it is not to be conducted out of the watershed, because the water balance in Lake Urriðavatn is to be maintained
- The stormwater is to be treated, because the water quality in the Lake Urriðavatn is to be affected as little as possible and may meet the requirements to the future water quality.
- The runoff intensity in developed areas should differ as little as possible from that before urbanization. The drainage system must be designed in such a way that damages and inconvenience during flood don’t occur
- The stormwater is to be utilized as a local resource for experience, recreation and biological diversity.

The strategy to achieve these goals and criteria are as follows:

- Developing guidelines for stormwater management, which each developer is required to follow it on his lot and property
- As much as possible of the stormwater should be conducted into the ground or conducted to natural runoff near the source, such as roof water, road water, water from parking places

- The treatment of stormwater is to be based on natural and not operation intensive solutions i.e. the methods of the nature itself such as infiltration, channels, dams, wetlands etc.
- Use source measurements to reduce supply of pollutants to the stormwater for example reduce fertilities, avoid use of building materials that contains environmental poison
- Separate clean stormwater i.e. roof water, from polluted stormwater to reduce the hydraulic loads on the treatment facilities. Stormwater from traffic areas should be treated before outlet into green areas or recipient.

These measures are relatively new in Nordic stormwater management, so the implementation may be a challenge.

POLLUTANT LOADS AND FUTURE CONCENTRATIONS IN LAKE URRIDAVATN

The land use will affect the future runoff the pollutant. Therefore calculating the production of pollutants, runoff of pollutants and effects in Lake Urriðavatn, the values in table 1 are used.

The calculations are based on standard concentrations as presented in StormTac, version 2005-04, StormTac (2005) and the areas distribution in table 1 and annual precipitation, 1035 mm/yr. In area U-1 are used values for commercial areas, otherwise values for row houses.

Runoff calculations are shown in table 3. The runoff coefficient is selected as traditional Nordic conditions, Lindholm et al: (2005).

Table 3. Area distribution and calculated runoff from developing areas to Urriðavatn, Aasen et al. (2005).

	Area (m^2) (A)	Runoff coefficient (C)	Intensity (l/m^2) (I)	Runoff (l/year) (Q)
Commercial area U-1	86000	0,90	1035	82 903 500
Other areas	408100	0,70	1035	295 668 450

StormTac is a watershed-based Excel model for; 1) the quantification of water flows and pollutant loads, 2) the design of stormwater treatment, transport and detention facilities, e.g. wet ponds, filter strips, wetlands, sewers, ditches, channels and detention basins. It is setting up water and mass balances and the quantification of acceptable loads and reduction needs for receiving waters, e.g. lakes.

In the calculations are expected some treatment of the stormwater by infiltration, in retention basins etc. The treatment efficiency, R, is based on continuous measurement in Norway that includes winter conditions and international experiences, Åstebøl et al. (2005).

Table 4. Efficiency in treating stormwater in urban areas in %, Åstebøl et al. (2005).

Pollutant	Treatment efficiency	Pollutant	Treatment efficiency
Phosphorus (P)	65	Chromium (Cr)	70
Nitrogen (N)	35	Nickel (Ni)	70
Lead (Pb)	70	Mercury (Hg)	70
Copper (CU)	70	Suspended solid (SS)	85
Zink (ZN)	70	Oil	80
Cadmium (Cd)	70	PAH	85

The calculations show the discharge of stormwater pollutants from urbanized areas after implementation of pollutant reducing measures and the resulting average concentration in Lake Urriðavatn. The calculations show therefore the expected increase in average concentrations in Lake Urriðavatn, exceeding today concentration after development.

The calculations of increase in concentration are based on a model for phosphorus loadings developed at Norwegian Institute of Water Research (NIVA), Berge (1987). The model is based on empiric data from a large number of shallow lakes. The model is also used to calculate the other pollutants except nitrogen. Metals and PAH have higher natural retention in lakes than phosphorus. The model will therefore overestimate the concentrations in the lake for those pollutants. For nitrogen there is only calculated dilution in the lake. Formula used to calculate the concentration in the lake is:

1) (P)i = (P)inn/Q where: (P)I = average concentration on P in the inlet(ugP/l)
(P)inn = annual discharge of phosphorous (kgP/year)

2) (P)$ = 0,436 x (P)I x $T_w^{-0.16}$ where: (P)$ = average concentration of phosphorous in the lake (ugP/l).
T_w = lag time (year) = V (m3)/Q (m3/year)

Table 5. Environmental limitations regarding heavy metals causing biological effects. Umhverfismálaráðuneyti Íslands (1999).

Metals (conc.)	**I (μg/l)**	**II (μg/l)**	**III (μg/l)**	**IV (μg/l)**	**V (μg/l)**
Copper (Cu)	<0.50	5-3	3-9	9-45	>45
Zink (Zn)	<5	5-20	20-60	60-300	>300
Cadmium (Cd)	<0.01	0.01-0.1	0.1-0.3	0.3-1.5	>1.5
Lead (Pb)	<0.2	0.2-1	1-3	3-15	>15
Chromium (Cr)	<0.3	0.3-5	5-15	15-75	>75
Nickel (Ni)	<0.7	0.7-15	15-45	45-225	>225
Arsenic (Ar)	<0.4	0.4-5	5-15	15-75	>75

By using StormTac the concentrations of some selected pollutants after the development were calculated. Table 4 shows the results from these calculations.

Table 6. Calculated total pollution to and concentrations of phosphorus (P), nitrogen (N), lead (Pb), Copper (Cu), Zink (Zn) and Cadmium (Cd) in the Lake Urriðavatn after the development, Aasen et al. (2005).

	P (ug/ℓ)	N (ug/ℓ)	Pb (ug/ℓ)	Cu (ug/ℓ)	Zn (ug/ℓ)	Cd (ug/ℓ)
Green areas	5391397500	97045155000	646967700	80879625	1347849375	16174193
Parking Imperv. surfaces	7623900000	83862900000	2287170000	3049560000	10673460000	34307550
Townhouses imperv. surfaces	70346237500	408008177500	4220774250	7034623750	23917720750	168830970
Total material produced	83361535000	588916232500	7154911950	10892893375	35939030125	219312713
- Retention	54184997750	206120681375	5008438365	7625025363	25157321088	153518899
Discharged	**29176537250**	**382795551125**	**2146473585**	**3267868013**	**10781709038**	**65793814**
From untouched areas	29372700000	528708600000	3524724000	4405905000	7343175000	88118100
=Total disch. to Urriðavatn	**58549237250**	**911504151125**	**5671197585**	**7673773013**	**18124884030**	**153911914**
Sum discharge to Urriðavatn	55.61	865.70	5.39	7.29	17.21	0.15
Calculated conc.	11.77	865.70	1.14	1.54	3.64	0.03

The calculated values for concentrations were then compared to the limitation values in table 3. It was shown that the values were within the requirements for the environmental limits for group II that belongs to Class B.

THE STORMWATER MANAGEMENT AT THE IKEA LOT

The IKEA lot is 17.4 hectares, where ca. 50% are impervious areas, roofs, 3.9 ha, and parking places, 4.9 ha. The IKEA building and the parallel BYKO building are located on the lava in the west part of the lot with good infiltration capacity, but the parking places are located on wetlands with low infiltration capacity. Buildings are located in east on rocky ground, with low infiltration capacity.

Stormwater on parking places is conducted to channels or grassed swales. Grassed swales allow for filtration of pollutants and recharge of groundwater. Channels and swales have filter materials of sand and a reactive filter material with adsorbing capacity on phosphorus and heavy metals. The topsoil is mixed with organic materials to encourage establishing vegetation. The filter masses should meet the requirements for infiltration and separation of particles and binding dissolved materials. The top grass shall prohibit clogging.

When the infiltration capacity in the channels exceeds during heavy rain, snowmelt etc. the surplus stormwater is conducted to collecting pipes, while surplus stormwater from the northern parking place runs to a wet detention basins for treatment and infiltration. The roof water from the IKEA buildings is infiltrated direct into the lava, while the roof water from the BYKO building goes into open channels for infiltration. The roof water from the buildings at the east side goes to wetlands or Lake Urriðavatn, Aasen et al. (2005).

The road water flows to the same facilities as the stormwater from the parking places. The selected stormwater solution will secure good water balance at the IKEA lot, while much of the stormwater is infiltrated into the ground as before urbanization.

DISCUSSION

Urban hydrology (UH) is a relatively new science within the much older general hydrology. Urban hydrology is the key discipline within stormwater management (SWM). There are still gaps in the knowledge and developments within both UH and SWM, even though the focus has been on UH and SWM during the last 30 – 40 years, while a little attention has been on UH and SWM in cold climate, Thorolfsson (2004). One of the reasons is the short come on urban runoff models and a detailed land use description: Furthermore urbanhydrological data collection with high resolution is still limited, even lacking, especially in cold climate. Another reason is the involvement of many disciplines such as engineers, environmentalists, technology, ecology, biology, limnology, policy and planning, geography and health studies etc.

Therefore some of the conclusions and decisions may be taken based on uncertain methods, data etc., but these are improving fast during last years. This study is also burden by uncertainty, but it was based on the best available knowledge at that time.

The study shows that it is possible to develop the watershed according to the plans without diminishing the water quality in Lake Urriðavatn.

RESULTS

The goal for Lake Urriðavatn is to keep it as untouched as possible, and to take care of water and land environment, and the natural ecology in the watershed.

The study shows that by use of Sustainable Urban Drainage Systems (SUDS), the water resources of Lake Urriðavatn will be acceptable, but the water quality Class B (a little touched) have to be accepted for some parameters.

The IKEA lot on the peninsula north of Lake Urriðavatn is finished implemented with SUDS solutions and it is functioning well.

The Urridaholt project won prizes for a walk-able community, with pedestrian-friendly streets and sustainable drainage systems, which is protecting the environmental quality of the watershed around Lake Urridavatn, ab – Arcitecture Boston (2008).

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SESSION 2: ARCTIC HYDROLOGY AND GLACIERS

CHANGES IN ICE REGIMES OF RIVERS IN THE NORTHERN PART OF EUROPEAN RUSSIA

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ABSTRACT

During the last 20-25 years ice regime on the rivers within the Northern part of European Russia was characterized by significant changes mainly due to a sustainable positive trend of winter air temperatures. The following ice regime characteristics have been considered: dates of ice-on and ice-off, ice cover duration and maximum ice cover thickness. Changes in ice regime characteristics have been estimated for 1980-2005 and compared with 1950-1979. Two methodological approaches have been used. It has been established that dates of ice-on on the mid-size and large rivers of the North-West part of European Russia were 2-6 days later if compared with the period 1950-1979. Besides, dates of ice-off appeared to be earlier in 2-5 days. This tendency would be kept during 2010-2015. Ice-on would be 2-10 days later and ice-off would be earlier, if compared with the norm. Mean ice cover duration in rivers in the last twenty five years became 2-10 days shorter, if compared with the previous 30-year period. In accordance with the forecast of further air temperature rise in winter and spring, the ice cover duration in the rivers of the Northern part of European Russia would be shorter during nearest 5-10 years. The maximum ice cover thickness on the rivers on the Northern part of European Russia during 1980-2000 was thinner by 2-7 cm on the average. Moreover, a positive tendency towards a thinner maximum ice cover thickness on the rivers of Karelia and Baltic Sea Drainage was observed for the whole 50-year period, whereas changes in the maximum ice cover thickness during 1950-1979 on the rivers of the North East of the European Russia were insignificant. It is possible to expect a development of those tendencies towards changes in the ice regimes in rivers which were established for the last twenty five years. The above changes in ice regimes on the rivers within the Northern part of European Russia require an adaptation of different economic branches related with water use in winter time.

INTRODUCTION

The problem of the assessment of ice regimes changing under the influence both the current and future changes of climatic situation is urgent for the rivers of the Northern part of European Russia. Many rivers within this territory are used intensively during winter time and such ice regime characteristics as dates of ice-on and ice-off, duration of ice events, and maximum ice cover thickness are very important and may frequently be characterized as a limiting factors. Dates and duration of ice events are associated with navigation and specific features of hydraulic structures construction in winter. The ice cover thickness is a determining factor to estimate the ice-bearing capacity and dates of ice-routes operation across rivers. Investigations show that ice regimes in the rivers within the Northern part of European Russia were changed significantly during the last 20-25 years. The main reason such a situation became a stable positive trend of winter air temperatures during this time.

STATE-OF-THE-ART OF THE PROBLEM

Ice regime characteristics observed during more than a hundred years on some rivers and lakes all over the world are one of the most reliable indicators of climate change. Therefore it is a great interest of scientific community towards the problem of their changing. The problems of rivers and lakes ice regime changes and their impacts on economics, ecology and vital activities are described in the publications of IPCC (2001), of the 27th Congress of International association of theoretical and applied limnology (Proceedings, 2000), of the International Association of Hydraulic Research (Proceedings, 2002).

In (Magnuson et al., 2000) ice events and ice-on durations have been analyzed for 39 water bodies (rivers and lakes) in the Northern Hemisphere for 1846-1995. It is shown that on the average, a later ice cover formation (by 5.7 days per 100 years) and earlier ice break-up (by 6.3 days per 100 years) were observed on those water bodies which corresponded to the air temperature rise by 1.2º C per 100 years on the average. Specific features of ice events on 16 rivers in the Arctic Russia are considered in (Vuglinsky, 2002). It was shown that during the last 20-25 years a strong negative trend in long term series of ice cover duration and maximum ice cover thickness took place for the largest Russian arctic rivers. In (Kuusisto & Elo, 2000), long-term series on ice regime characteristics for rivers and lakes in northern Europe have been analyzed for the period from the start of observations till 1990. It is noted that during the last 10 years a shorter ice cover duration and thinner maximum ice cover thickness on water bodies are due to warm winters. During 1981-2001 mean ice cover thickness on the rivers of the Belarus became about two times thinner; concurrently, the duration of ice events decreased by 5-15% and the duration of ice-on decreased by 15-30% (Skuratovich, Komarovskaya, Chekan, 2004).

The problem of the assessment of climate change effect on ice regimes in water bodies was emphasized at the VI All-Russian Hydrological Congress (St.Petersburg, 28.09-01.10.2005). For example, on the rivers within the Maritime Territory significant negative tendencies in the ice cover duration was discovered (Makagonova, 2004). A negative trend in ice cover thickness were discovered for some Polish and Russian lakes. In Finland, both decreasing and increasing trends can be found in the maximum ice cover thickness time series (Assessment, 2008).

METHODOLOGICAL APPROACHES

Analysis of ice events dynamics in the rivers of the Northern part of European Russia has been made on the basis of observation data series on ice regime characteristics and ice cover thickness published in hydrological yearbooks. Data from 45 hydrological stations installed on large and mid-size rivers within the study area have been used for the analysis. An electronic database has been prepared for the period from the beginning of observations up to 2000-2005 (Table 1).

Table 1. List of selected stations used for analyses.

River - station	Drainage area, sq.km	Periods of observation	
		Maximum ice cover thickness, cm	Ice events
Vodla river - Kharlovskaya	12000	1954-2005	1937-2005
Volkhov - Volkhovo	70000	1954-2005	1946-2005
Shelon - Zapolye	6820	1954-2005	1945-2005
Luga – Tolmachevo	6350	1954-2005	1945-2005
Luga - Kingisepp	12800	1954-2005	1945-2005
Mezen - Malonisogorskaya	56400	1953-2002	1932-2002
Pechora – Ust-Tsylma	248000	1953-2002	1914-2002
Severnaya Dvina - Ust-Pinega	348000	1953-2002	1882-2002
Onega - Porog	55700	1953-2002	1918-2002

Assessment of ice regime changes has been made for 1980-2005 in comparison with 1950-1979. The period 1950-1079 was selected as a base period because it was characterized by missing trend in air temperature series, whereas mean annual air temperature for that period was close to norm. As there were gaps in data, low quality of the data and a necessity to use additional data characterizing winter temperature background adjacent to the study river basins, two methodological approaches have been applied.

The first approach is based on a comparison of averaged characteristics of ice regimes for 1950-1979 and 1980-2005 to determine the gradients of these changes and use this gradient for future assessments. The second approach is based on the establishment of correlation between ice regime characteristics and winter air temperatures for the study periods, with its future use with the account of prognostic values of winter air temperature. In most cases both of these approaches were used for the analyses.

ASSESSMENT OF CHANGES IN ICE REGIME CHARACTERISTICS

During the last 20-25 years, earlier ice cover formation (by 2-6 days on the average) was observed on large rivers in the North of the European Russia, if compared with the period 1950-1979; as a whole, it corresponded to a tendency towards a little fall of mean air temperature in autumn.

In the rivers of the North West of European Russia a stable positive tendency was observed during the 55-year period (1950-2005) on the dynamics of the dates of the start of ice-on and a negative tendency – in the dynamics of the dates of ice-off, i.e. later dates of ice-on and earlier dates of ice-off. If compared with the base period 1950-1979, this shift during the last twenty five years equaled 5-7 days on the average. This tendency in these regions would be kept during 2010-2015.

Mean ice cover duration in the last twenty five years in middle and large rivers of the Northern part of European Russia became 2-10 days shorter, if compared with the previous 30-year period (Figure 1). Moreover, there is an evident tendency towards a shorter ice cover duration by 25-30 days on the average observes in the rivers of the North West of European Russia during the whole 55-year period (Fig.1); meanwhile in the rivers of North East of the European Russia the ice cover duration was practically the same as it took place in 1950-1979.

In accordance with the forecast of further air temperature rise in winter and spring, the ice cover duration in the rivers of the Northern part of European Russia would be shorter. In some rivers (Pechora, Northern Dvina, and Onega) this shortening in 2015 may be 10-15 days.

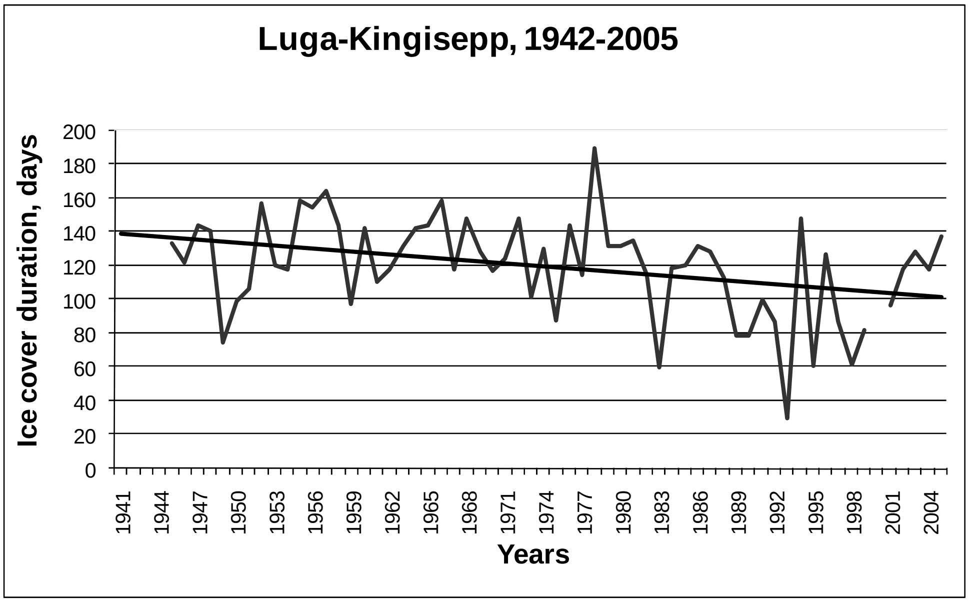

Figure 1. Changes in ice cover durations for the Luga river during 1942-2005 (North West of European Russia).

Practically in all mid-size and large study rivers in the Northern part of European Russia a decrease of maximum ice cover thickness by 2-7 cm on the average was observed during the last twenty five years. If compared with the previous 30-year period. Moreover, in many rivers, a negative tendency towards a thinner maximum ice cover thickness was observed during the whole 50-year period (Fig.2); on the contrary, in the river reaches of the Pechora the maximum ice cover thickness became a bit thicker.

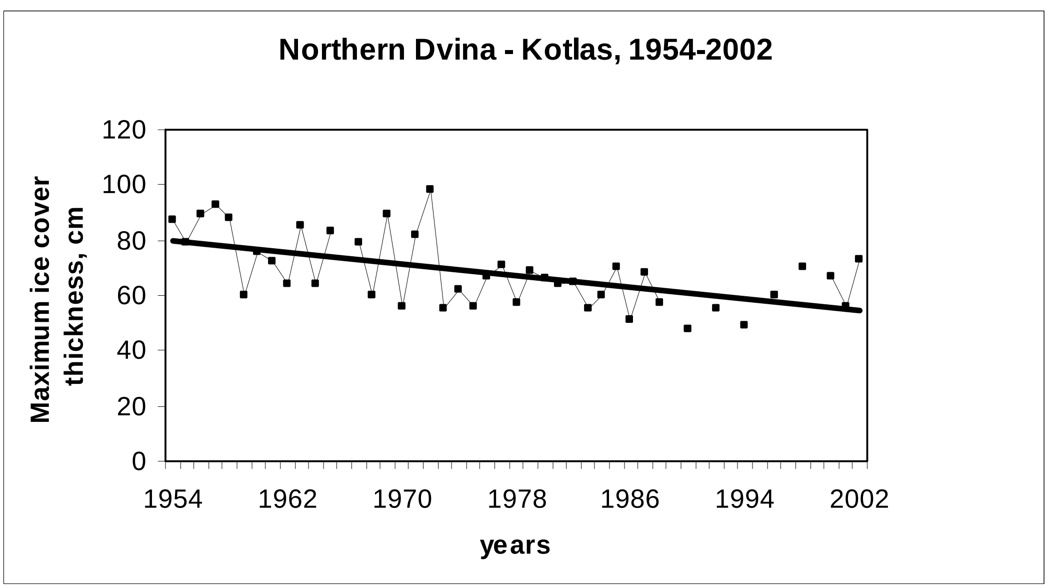

Figure. 2 Changes in the maximum ice cover thickness in the Northern Dvina river for the period 1954-2002 (North East of the European Russia).

In accordance with a tendency towards higher spring and winter air temperatures, a further thinning of the maximum ice cover thickness in mid-size and large rivers of the Northern part of European Russia is expected. This thinning by 2010-2015 would be small (if compared with the present tine) and would equal 3-7 cm.

Combined data on the changes in ice cover duration and maximum ice cover thickness in large rivers of Russia are given in Table 2.

Table 2. Changes in mean ice cover duration and maximum ice cover thickness in large rivers of the Northern part of European Russia in 1980-2005 relative to 1950-1979.

River	Changes in mean ice cover duration, days	Changes in mean maximum ice cover thickness, cm
Volkhov	0-2	-1-2
Luga	-2-4	-2-4
Northern Dvina	-4-6	-2-5
Onega	-2-6	-1-3
Pechora	-2-4	-2-6

As further climate warming is most probable for the next decades in the territory of Russia, positive trends of winter air temperatures in particular, it is possible to expect a development of those tendencies towards changes in the ice regimes in rivers which were established for the last twenty five years. Therefore, it is very important to adapt different branches of economy to the changes in ice river regime which may cause both positive and negative results.

CONCLUSION

The above changes in ice regimes on the rivers within the Northern part of European Russia require an adaptation of different economic branches related with water use in winter time. If the discovered tendencies remain in the ice regime variations on the rivers of the European Russia, two major branches of economics (navigation and road transport) would be affected most seriously. Moreover, adaptation measures may be different. For example, a longer period of navigation due to shorter periods of ice-on would increase freight turn-over of the river fleet on the one hand, and, on the other hand, it would hinder the arrangement of ice crossings across the river thus causing great difficulties in operation of the freight traffic and passenger transport across these rivers. This situation is typical of many northern towns along the river banks, e.g. Arkhangelsk on the Northern Dvina.

The observed thinning of the maximum ice cover thickness on the rivers within the European Russia leads to a lower carrying capacity of ice routes for the traffic, and consequently, to less volumes of the transported cargoes at the same frequency of vehicles motion. A lower efficiency of ice roads operation is also connected with a shorter ice-on period. Under such conditions adaptation measures are required, aimed at optimization of traffic along the ice roads. These measures may be as follows: more intensive transport along ice routes, artificial ice increment on the reaches with critical ice cover thickness, arrangement of combined routes along ice and, etc. Adaptation measures will be also required for planning the ice crossings across rivers.

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ARCTIC HYDRA: THE ARCTIC HYDROLOGICAL CYCLE MONITORING AND ASSESSMENT PROGRAM

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ABSTRACT

The scientific goals of the Arctic-HYDRA project are: To characterize variability in the Arctic Hydrological Cycle (AHC) and examine linkages between atmospheric forcing and continental discharge to the ocean; to assess the historical response of the Arctic Ocean to variations in freshwater input from rivers and net precipitation over the ocean; to attribute to specific elements of the AHC or to external forcing the sources of observed spatial-temporal variability in the land-ocean-ice-atmosphere system; to detect emerging changes in the contemporary state of the AHC in near real time and to place such changes into a broader historical context. Starting during the International Polar Year (IPY), Arctic-HYDRA also forms part of the parallel longer term (10-15 yr) objectives of the ICARPII WG7 project "Terrestrial Cryospheric & Hydrologic Processes and Systems".

INTRODUCTION

Scientists focussing on issues related to the hydrological cycle in the Arctic have convened several times since 2005 to create and develop plans for a new international research consortium, Arctic-HYDRA. The main overarching questions that this consortium will seek to answer in future studies are:

- What is the role of the unified Arctic Hydrological Cycle (AHC) in the global climate system?
- What are the feedbacks of changes in the Arctic Hydrological Cycle on the regional and global climate systems?
- What are the impacts of changes in the Arctic Hydrological Cycle on the biology, biogeochemistry, and human society?

The plans for Arctic-HYDRA grew out of the need to link various ideas for Arctic research projects that were submitted to the planning office for the International Polar Year (IPY). It is recognized that these questions will not be definitively settled in the time span of the IPY (2007-2009), and hence they are meant to "set the stage" for the Arctic-HYDRA in the context of other important Arctic science initiatives. The Arctic-HYDRA planning exercise entrained the perspectives of these ongoing or planned studies, as well as the individuals who have contributed to these intellectual frameworks. The Arctic-HYDRA community is in a unique position to ensure that its science agenda, to the highest degree possible, will build on existing questions and those that are cast to complement those of other ongoing projects.

THE ARCTIC IN THE CLIMATE SYSTEM

The Arctic exerts important influence over global climate through feedback mechanisms, by which Arctic processes can amplify initial climate changes (Serreze and Barry, 2005). Examples of such feedbacks are the warming effects of lowered surface reflectivity accompanying such changes as the decrease in snow and ice cover and the northward expansion of forests into tundras. Another important effect is potentially due to increased precipitation and runoff into the Arctic Ocean from adjacent continental areas, which would exert negative influence on deep-water formation and reduce sea ice formation. The resulting slowdown in the worldwide thermohaline circulation would slow the transport of carbon dioxide to the deep ocean, thereby allowing more rapid buildup of CO_2 in the atmosphere. A third important effect is the influence of climatic warming on the exchange of greenhouse gases between the atmosphere and Arctic soils and sediments, where large amounts of carbon are stored (ACIA, 2005).

Arctic HYDRA represents a new, consortium-based international study that aims to provide a quantitative picture of the state of pan-Arctic hydrological system in the years of the International Polar Year campaign and beyond. It will represent a critical benchmark against which to assess future change, with focus on cold land processes, on environmental impacts, the greening of the Arctic, on permafrost and frozen soils, on ecosystems, on changes and variability, aerosols, environmental change and people.

SOCIETAL ISSUES

Water is a fundamental component linking many of the environmental changes in the Arctic region, and society demands answers to how a changing Arctic Hydrological Cycle impacts humans, ecosystems, and earth systems. Specific societal issues of concern include commerce, shipping and resource extraction, pollution, coastal erosion, fisheries, forestry and vegetation changes. The Arctic Climate Impact Assessment (ACIA, 2005) summarized many of these changes, and stressed that not only are many of these already taking place, but they are expected to accelerate over the next 100 years and beyond. Furthermore, Working Group II of the IPCC Fourth Assessment Report stated with high confidence that climate change (e.g. temperature increases) is affecting natural systems, including changes in snow, ice, and frozen ground/permafrost (IPCC, 2007). There is also high confidence that impacts will include increasing coastal erosion, increasing seasonal permafrost thawing depth, and reduced extent of permafrost and sea ice. While there are both projected benefits and negative impacts of these changes, there is increasing concern about the extent to which the inhabitants of Arctic regions will be able to adapt to these changes in the future.

OBSERVATIONAL STATUS

For process studies and modelling, the Arctic HYDRA consortium will build on existing meteorological, hydrological and cryospheric data sets from the Arctic region. The project will also facilitate the development of both research and sustained observations in the Arctic.

Temperature records, glacier and permafrost observations and data on sea ice extent and thickness all provide strong evidence for Arctic warming in the past decades (ACIA, 2005). A warming of 2-3 °C has been recorded in most of Alaska, in NW Canada and parts of Siberia. On average, annual Arctic temperature has increased by almost twice the rate as that of the rest of the world, with some variations across the region. Precipitation in the Arctic has, on average, increased by 8% in the past century.

Widespread cryospheric changes throughout the Arctic region are well documented. Most Arctic glaciers and ice caps have been in decline since the early 1960s, with this trend speeding up in the 1990s. Less certainty exists about recent changes in the state of the Greenland ice sheet, but the most recent assessments indicate accelerating mass loss from the ice sheet during the 1990s up to 2005. According to the latest IPCC Assessment report, Greenland was in 1996 losing about 96 km^3 per year in mass from its ice sheet. In 2005, this had increased to about 220 km^3 a year due to rapid thinning near the coast, while in 2006 it was estimated at 239 km^3 per year.

The Arctic sea ice cover has been in dramatic decline over the past four decades. Between 1974 and 2003 the annual average sea ice extent decreased by 8% and average arctic-wide sea ice thickness decreased by 10-15%. Recent observations (NSIDC, 2007) indicate that sea ice retreat is more rapid than estimated by any of the computer models used by IPCC in preparing its 2007 assessment.

About 70% of the total terrestrial Arctic drainage area of $22.4*10^6$ km^2 is monitored at present. The mean annual drainage is estimated to be 3200 km^3 and the four largest drainage basins, the Ob, Yenisey, Lena and MacKenzie, contribute about 68% of the total gauged volume discharge to the Arctic Ocean.

THE PROBLEM OF DECLINING NETWORKS AND THE NEED FOR A PAN-ARCTIC APPROACH

Access to comprehensive and reliable data sets on Arctic hydrology is of crucial importance for studies focusing on the role of the Arctic in the climate system. The rivers draining to the Arctic Ocean redistribute moisture from temperate regions to the high latitudes, and also connect very large and heterogeneous areas to the Arctic Ocean and its shelf seas.

The Arctic Runoff Data Base, ARDB, which is collected and maintained at the Global Runoff Data Centre, GRDC, in Koblenz, Germany, provides station information and data series of runoff gauging stations in the Arctic region (GRDC, 2006). With more than 2400 stations represented, this is the most complete international dataset on daily (1024) and monthly (2193) runoff in the Arctic. Another service, containing freely available Pan-Arctic river runoff data, is compiled at the University of New Hampshire, USA. The University of New Hampshire maintains a service providing near-real-time river discharge data for nearly 60 stations (ArcticRIMS, 2007).

Although Arctic nations currently expend about $100 million annually on the collection of hydrologic data, the number of sites at which data are collected is declining (Shiklomanov and Shiklomanov, 2003). Many sites with long-term records have been discontinued in the past several years as funding has failed to keep pace with costs. As development and resource extraction in the Arctic increases, the need for hydrologic data also increases. The hydrological station network is still very uneven and no observations are carried out in one-third of the region. The observation series are of unequal lengths and numerous gaps exist in several records. Recent studies have highlighted the fact that the decline in station density has been greatest in regions that are projected to undergo the largest increase in temperature during the 21st century; i.e. in areas adjacent to the Barents, Laptev and Kara seas (Bring et al., 2007).

Arctic-HYDRA will provide a framework to:

- Design and execute a systematic process to evaluate the utility of key historical and operational data sets depicting the geospatial distribution of water cycle variables, using biogeophysical information generated over the satellite era and capable of assessing the fully pan-Arctic domain.
- Apply the evaluation process to create optimal deployments of monitoring network and remote sensing resources to construct an operational, contemporary depiction of the pan-Arctic water cycle during IPY and beyond.
- In concert with regional and global modelers, assess the effectiveness of current monitoring resources to detect plausible scenarios of Arctic environmental change.

- To provide feedback to data providers on the efficacy and value of their data sets to the user communities and to propose concrete suggestions on ensuring their relevancy.

TOWARDS AN INTEGRATED SYSTEM FOR ARCTIC HYDROLOGY STUDIES

Cast as a set of scientific objectives to support the overarching questions listed in the introduction, Arctic-HYDRA seeks to:

- Characterize variability in the Arctic Hydrological Cycle (AHC) and to examine linkages between atmospheric forcing and continental discharge to the ocean.
- Assess the historical response of the Arctic Ocean to variations in freshwater input from rivers and net precipitation over the ocean.
- Attribute sources of observed spatial-temporal variability in the land-ocean-ice-atmosphere system to specific elements of the AHC or to external forcing.
- Detect emerging changes in the contemporary state of the AHC in near real time and to place such changes into a broader historical context.

Given the scope of these objectives and the relatively short time-frame of the IPY, Arctic-HYDRA also forms part of the parallel longer term (10-15 yr) objectives of the ICARP-II (International Conference on Arctic Research Planning) Working Group 7 (WG7) project, "Terrestrial Cryospheric & Hydrologic Processes and Systems". The Arctic-HYDRA project will use IPY as a stepping stone to longer-term, comprehensive water cycle studies (e.g. ICARP-II) and within ISAC (International Study of Arctic Change).

The Arctic-HYDRA project will consist of a core network of observations of the Arctic Hydrological Cycle (Arctic-HYCOS), coupled with a suite of intensive, focused process studies that are based on in-depth measurements and modeling of the individual components of the Arctic Hydrological Cycle (AHC). Furthermore, hydrological models and data assimilation techniques will be developed to generate a comprehensive, integrated description of the AHC including the feedbacks between the atmosphere, cryosphere and the oceans. The project will have a data management and information system in accordance with IPY and WMO protocol. It will establish links with other relevant clusters, e.g., meteorology, climatology, cryosphere, including permafrost, snow-cover and glaciers, biosphere and societal issues affected by the AHC.

CONCLUDING REMARKS

Arctic-HYDRA represents a unique opportunity to advance the science of the Arctic Hydrological Cycle and to establish a legacy of novel and comprehensive pan-Arctic observational networks that will contribute to global earth observing systems. The project has been endorsed by the ICSU/WMO Joint Committee for the International Polar Year 2007-2008, and the Committee has stated that the proposal 'includes very strong scientific, education and outreach components and demonstrates a high level of adherence to IPY themes and goals', and that the activity would represent a 'prominent and valued part of the IPY program'. Arctic-HYDRA is supported by WMO Hydrology and Water Resources (HWR) Programme, WMO Commission of Hydrology (CHy), WMO Commission for Basic Systems (CBS), WCRP/CliC and the Global Runoff Data Centre (GRDC), and is funded by the Nordic Council of Ministers as an IPY activity. The effort includes participation of representatives from all Arctic countries and Japan, as well as participation of all Arctic National Hydrological Services. Further information is available at: http://arcticportal.org/arctichydra

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SPATIAL PATTERNS OF DECLINE IN PAN-ARCTIC HYDROLOGICAL MONITORING NETWORKS: A VULNERABILITY MAP

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ABSTRACT

The last decades of observed rapid and significant changes to the Arctic hydrological system indicate an ongoing transition to a state not previously observed in recent history, which stresses the need for hydrological and hydrochemical observation networks that are adequate for detecting, understanding and modeling these changes. Recent studies have reported a widespread decline in these networks, but little information is available on where the decline has been most critical, and how it relates to the distribution of socio-economic and climatic pressures on water resources in the pan-Arctic drainage basin. We present a quantitative picture of the spatial patterns of decline in Arctic hydrological monitoring networks. We also analyze which Arctic drainage basins that are left most vulnerable by this decline, due to their combination with socio-economic and climate pressures. Results indicate that for basins where the hydrological monitoring decline has been higher than average, population density and economic production intensity are also frequently above average. Furthermore, diverging spatial patterns in future modeled and recently observed temperature trends makes it difficult to determine the real vulnerability of these basins to temperature change pressures.

INTRODUCTION

During the last few decades, significant and rapid changes to various components of the pan-Arctic hydrologic cycle have been observed (e.g., Serreze et al., 2000; Houghton et al., 2001; Peterson et al., 2002; Stroeve et al., 2005). In order to detect, understand and adapt to these changes, a well-distributed hydrological monitoring network with openly accessible data of high quality is vital. Unfortunately, several studies have shown a marked decline in hydrological monitoring networks since the period of their maximum extent in the 1970s (Lammers et al., 2001; Shiklomanov et al., 2002; Hinzman et al.; 2005, Walsh et al.; 2005). However, little information is available on where the decline has been most critical, and how it relates to the distribution of socio-economic and climatic pressures on water resources in the pan-Arctic drainage basin.

This paper presents a quantitative picture of the spatial patterns of decline in accessible data from Arctic hydrological monitoring networks. It also analyzes which Arctic drainage basins that are left most vulnerable by this decline, due to their combination with socio-economic and climate pressures.

METHOD

In order to analyze the spatial distribution of decline in accessible hydrological monitoring data, we introduce a measure to represent this decline in a coherent way across drainage basins with varying sizes and station densities. For this purpose, we calculate first the development of Arctic Runoff Database (ARDB) stations with accessible discharge data over time, from 1950 to present. This development analysis shows that the period 1975-79 represents the period of peak accessible monitoring data density. As a general measure of relative network density decline, we therefore use the ratio between the number of ARDB stations with accessible discharge data for the relatively recent period of 1995-99 with the corresponding number of stations for the peak-monitoring period of 1975-79. Since we cannot distinguish between stations that have actually been closed and stations that have ceased to provide access to data, the term network density in this context represents the density of stations with accessible data.

Using this measure, we develop a vulnerability map for basins with above-average decline in network density, based on calculated basin-by-basin population density, economic production intensity and projected future temperature change. Basins with above-average values for these parameters, relative to the average value for the entire pan-Arctic drainage basin, are identified, and an overlay map is created, indicating how many of the parameter values that are above average for each basin. Economic production data are based on the G-Econ (Geographically based Economic data) database (Nordhaus et al., 2006). Population and population density are calculated from the Gridded Population of the World database (CIESIN, 2005). Drainage basins are defined using the 30 × 30-minute STN-30p drainage network (Simulated Topological Network; Vörösmarty et al., 2000).

Furthermore, observed present and modeled future temperatures for a range of aggregated Arctic drainage basins are also analysed in relation to the relative network decline. Recent observed temperature trends for 1995-2002 are obtained by combining gridded data sets CRU CL 1.0 (New et al, 1999) and CRU TS 2.1 (Mitchell and Jones, 2005). Future temperature change fields for the period 2040-2069, relative to 1961-90, are acquired from the IPCC Data Center (IPCC, 2007) and averaged for three Global Circulation Models

(CGCM2, ECHAM4 and CCSR/NIES) and the SRES (Special Report on Emission Scenarios) A2a scenario. Furthermore, the range in value between the three different Global Circulation Models is calculated as a measure of the degree of uncertainty in temperature change predictions.

RESULTS

Fig. 1 shows an overview of the pattern of monitoring network decline in the pan-Arctic region. Overall, the number of stations with accessible data declined from 1877 for the period 1975-1979 to 837 for the period 1995-1999, corresponding to an average remaining density ratio of 0.45. In 53 basins, ranging in size from 800 to 373,000 km^2, the ratio was zero, indicating that there were no stations from the 1970s that provided data also for the 1990s. Many of these basins, however, were small and often contained only a single station that previously provided data from the basin. Six basins had a ratio greater than one, and in all cases this corresponded to the addition of data from one station in the basin.

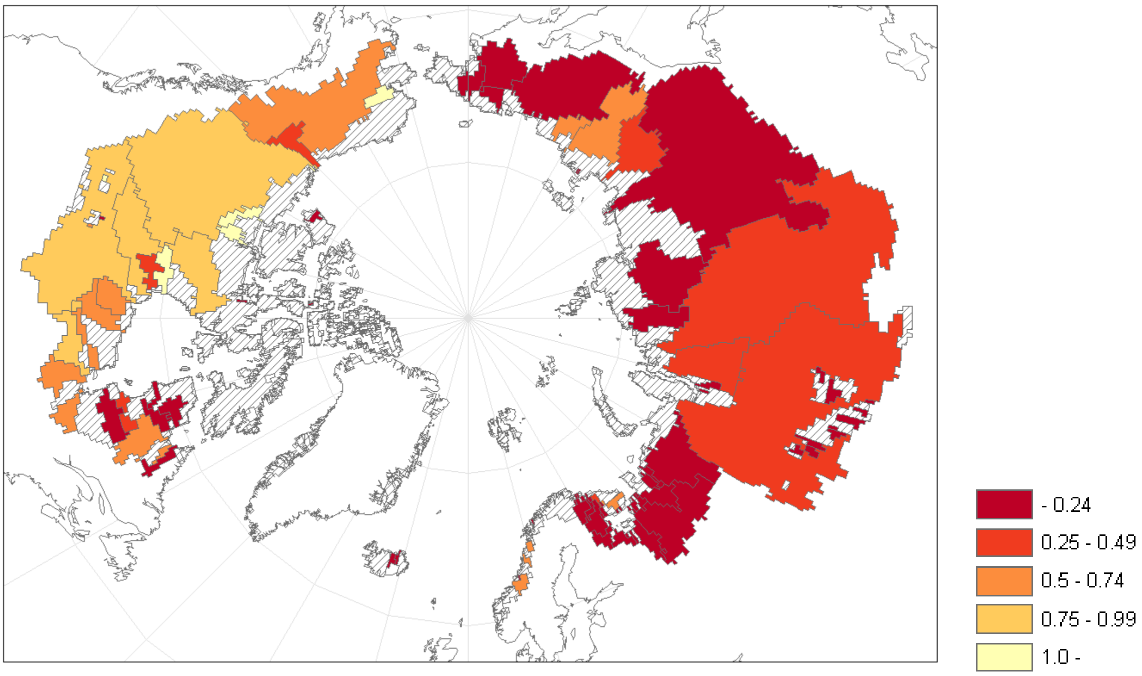

Figure 1. Evolution of accessible discharge monitoring station density in monitored Arctic drainage basins, expressed in the ratio of stations with accessible data for the basin from the period 1995-1999, compared to the period 1975-1979.

In Fig. 2, we show a vulnerability map for basins with above-average decline in monitoring networks, based on an overlay of basins with above-average values for population density, economic production intensity and projected future temperature increase. A value of 1 for a given basin means that of the four parameters, only the network decline was above average.

Values 2-4 indicate that one to three of the other parameters were also above the average.

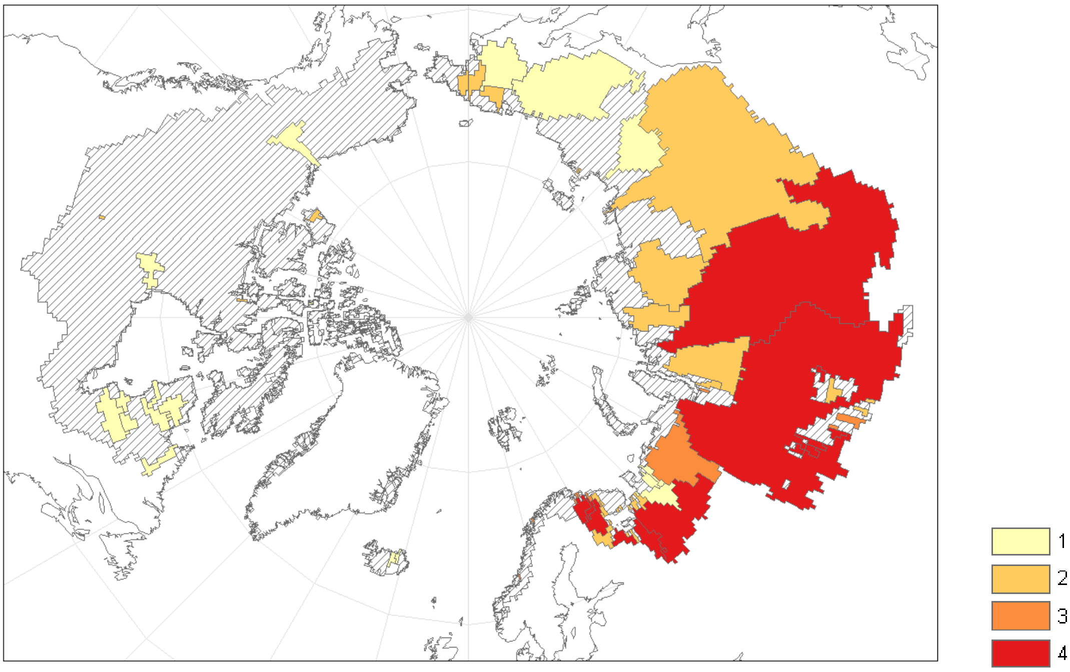

Figure 2. Vulnerability map for pan-Arctic drainage basins with above-average decline in accessible discharge monitoring station density. A value of 1 indicates that network decline is above average for the basin, while higher values indicate how many of the parameters population density, economic production intensity and predicted future temperature change that are also above average.

In total, these basins with above-average decline and pressure vulnerability comprise 50% of the whole pan-Arctic drainage basin, and 64% of the area of all the monitored basins within it. Furthermore, 81% of the population in the whole pan-Arctic drainage basin resides in basins with above-average decline in network density, and out of this share, 95% lives in those basins where also the population density is above average. The pattern for economic production is similar, with basins that have above-average decline in networks comprising 73% of the total economic production, 91% of which is in production-intense areas.

In Fig. 3 we show also how the decline in discharge monitoring relates to recently observed and projected future temperature trends. Four Eurasian drainage basins – draining into the Barents, East Siberian, Laptev, and Kara Seas – have been subject to the greatest absolute and relative decrease in discharge monitoring density among the stations in the ARDB database (Fig. 3). The difference between the number of stations with data accessible for the two investigated periods are -243, -63, -134, and -520 stations for these four drainage basins, respectively. With respect to observed climate trends during the period 1995-2002, these four drainage basins with the most significantly decreasing monitoring density have so far experienced the smallest observed temperature change (Fig. 3, left). However, the pattern is reversed when regarding projected future temperature trends. The four drainage basins with the most pronounced decline in discharge monitoring density are then the ones that are expected to have the greatest future temperature change (Fig. 4, right). The absolute range, i.e., the uncertainty of modeled future temperatures, is also particularly large for these four areas.

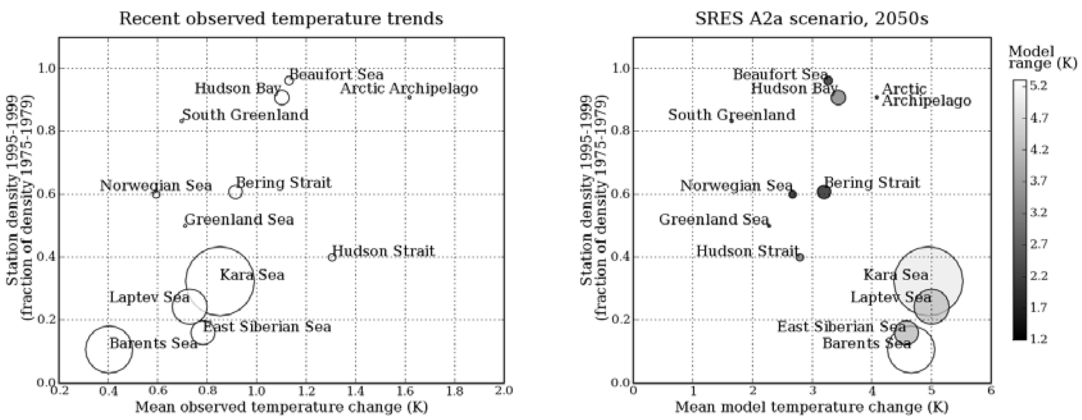

Figure 3. Evolution of accessible discharge monitoring station density in Arctic drainage basins, from the period 1975-1979 to the period 1995-1999 (vertical axis), in relation to the observed temperature increase in 1995-2002 compared to 1961-1990 (horizontal axis; left) and the GCM-projected temperature change fields for the 2050s (horizontal axis; right). Circle sizes correspond to the absolute reduction in number of discharge monitoring stations with accessible data. On right, circle colors represent the range in mean projected temperature by the three different GCMs (CGCM2, ECHAM and CCSR/NIES).

DISCUSSION AND CONCLUSIONS

If the decline of monitoring density had been evenly distributed across monitored regions, we should expect a relative measure of network decline to exhibit small variation between basins. However, we show that network decline is concentrated to certain areas of the pan-Arctic. Furthermore, for the basins where the hydrological monitoring decline has been higher than average, population density and economic production intensity are also frequently above average. These basins also contain the largest fraction of the total population and economic production in the pan-Arctic.

There are further diverging spatial patterns in the hydrologic monitoring decline combination with the projected future and the recently observed temperatures. This divergence implies that the basins with the greatest decline in discharge monitoring density have so far experienced the smallest observed temperature change, while climate change modelling indicates that they may experience the greatest temperature change in the future. This divergence between currently observed and projected future temperature changes makes it difficult to determine the real vulnerability of these basins to temperature change pressures.

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POTENTIAL IMPACT OF CLIMATE CHANGE ON THE SNOWPACK IN A PARTLY-GLACIATED BASIN IN CENTRAL SWITZERLAND

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ABSTRACT

In the Alps, hydropower companies are concerned about how climate change will affect runoff from partly glaciated and snow-rich catchments. Therefore, we modelled the snowpack response to global warming in a high alpine basin in central Switzerland. The snow cover and glacier was simulated using Alpine3D, a detailed model of mountain surface processes. To model snow conditions expected at the end of this century, we modified temperature, precipitation and longwave radiation measured from 1996 to 2001 using six regional climate model projections, for two different emission scenarios of greenhouse gases. According to our simulations, we can expect the snow season to shorten with up to one month at the beginning of the winter, and with at most two months at the end of the season compared to today. The seasonal peak snow water equivalent, averaged for the whole basin, decreases by 22%. The difference in the response between the emission scenarios is rather small. Most marked are the effects on higher altitudes, which will also completely melt-out during summer and where no snow remains for glacier accumulation.

INTRODUCTION

Today, snow and glacier melt is the single most important contribution to runoff throughout the year in alpine catchments (Swift et al., 2005). The accumulated melt water is commonly stored in reservoirs and is important for hydropower production in Switzerland (Schaefli et al., 2007). In the Alps, the amount and timing of basin runoff will likely change with a warming climate because of earlier snow cover depletion and glacier retreat (Braun et al., 2000, Beniston et al., 2003b). The most pronounced effects are awaited for glaciated catchments, where high runoff levels shifts from summer to spring because melt rates of the ice sheet decrease (Hagg et al., 2007) after a transitional phase of increased runoff (Ye et al., 2003). Along with an expected decrease in melt water contribution to runoff, higher evaporation rates in summer are awaited to affect the runoff as well (Calanca et al., 2006). As evaporation losses are greater from surfaces free from snow and ice, they largely depend on how long the snow season lasts, and by which grade the basin is glaciated. From the perspective of water resource management, we need to estimate how the snowpack will respond to climate change to make better predictions of future runoff characteristics in high alpine catchments, especially in basins with a complete glacier loss.

Studies on snow measurements in the Swiss Alps show that the snowpack duration depends on elevation and typical weather conditions prevailing during individual winters, and that it is sensitive to global warming (Laternser and Schneebeli, 2003, Scherrer and Appenzeller, 2006). Using predicted changes in temperature and precipitation, and by evaluating historic snow measurements the snow season in the Alps is expected to shorten with at most two months at the end of this century (Beniston et al., 2003a, Wielke et al., 2004). Snow model simulations show similar trends, with lower accumulated snow volumes throughout the winter and earlier melt-out dates, using temperature index methods (Jasper et al., 2004), or energy balance models (Beniston et al., 2003b, Keller et al., 2005). However, these studies typically focus on larger catchments at lower altitudes and have a limited representation of snow processes especially in complex terrain: The models represent the snow cover on a single layer basis, and therefore with a rather simple representation of the energy budget.

Here, we examine how the snow cover in a high alpine and partly glaciated basin in central Switzerland responds to climate change. The snowpack and glacier was simulated with the physically based and spatially distributed model Alpine3D describing snow cover and mountain surface processes (Lehning et al., 2006). Meteorological measurements from 1996 to 2001 were considered to represent the present conditions well, because the particular period covers winters with both shallow (1998) and thick snowpacks (1999). Longer periods could not be modelled due to high computation times. The observed data series was modified to simulate the snow cover characteristics expected for the end of this century using six different regional climate model (RCM) projections.

MATERIALS AND METHODS

Location

The Damma glacier basin in central Switzerland (46˚38.1´N, 8˚27.6´E) has an area of $12km^2$, and elevation ranges from 1844 to 3621m a.s.l. The runoff of this area discharges entirely to a hydropower dam (Göscheneralpsee). Measurements of the glacier extension started in 1921, and since then the glacier front has retreated around 10m per year (Hammerli et al., 2007). A last push forward of the glacier was recorded between 1972 and 1991. Today, the glacier occupies about $5km^2$ of the total basin area. The catchment is underlain by granite bedrock, and glacier tills are to be found on the forefield covered with low vegetation.

Model description

We use the distributed model Alpine3D, which is designed for high-resolution simulations of mountain surface processes in alpine terrain (Lehning et al., 2006). The snowpack is modelled by solving the heat and mass transfer equations of the snow cover, and the thermal conductivity and viscosity within each layer of the snowpack are determined by simulations of the snow microstructure and its rate of change (Lehning et al., 2002). The snow module as a stand-alone application (SNOWPACK) has previously been used for climate change research (Rasmus et al., 2004), and is also capable to represent phase changes and densification from snow to ice. Therefore it can be used for simulating glacier development (Obleitner and Lehning, 2004, Michelmayr et al., In review). The radiation distribution in the basin is described by a complex terrain radiation module (Fierz et al., 2003, Landl and Lehning, In review), which uses a view-factor approach to model the influence of the surrounding terrain (Oke, 2000). The runoff module is based on a conceptual approach and needs to be calibrated against runoff measurements from the specific catchment (Zappa et al., 2003). As runoff has not been measured in the Damma glacier basin before 2007, we concentrate on the snow cover response to climate change in this paper.

Alpine3D is initialized with a digital elevation model (100m resolution) and a description of the land use on this grid. Further, the model is forced by hourly meteorological measurements or predictions of air temperature, relative humidity, wind speed, precipitation and incoming long- and shortwave radiation. The first four parameters can be spatially interpolated to the grid by geostatistical models from several surrounding meteorological measurement stations.

Description of simulations

Three model simulations were completed for this study. The present conditions in the basin were simulated using meteorological measurements from 1996 to 2001, and the model was initialized with the glacier extent of year 2000. To analyse the seasonal snow cover characteristics, we removed all glacier ice and snow older than one year, so-called firn snow, from the complete ice and snow profiles. The effect of climate change on the snowpack was simulated separately for the two different emission scenarios A2 and B2 of greenhouse gases defined by the Intergovernmental Panel on Climate Change IPCC (Nakicenovic et al., 1998). The scenario A2 predicts a strong increase of carbon dioxide, whereas the scenario B2 describes moderate raises of greenhouse gases. By that time, we expect the glacier to be completely vanished, independent of assumed raises of carbon dioxide (Horton et al., 2006, Zemp et al., 2006).

Model input

As model input we used data from twelve stations surrounding the Damma glacier basin. These stations are located at a distance of 10 to 41km away from the study area, and represent an altitude range from 451 to 3580m a.s.l. Temperature, humidity, wind speed and precipitation input data were distributed to the grid using WINMET (Zappa et al., 2005), a tool for geo-statistic interpolation of meteorological data. As interpolation method we selected inverse distance weighting (Shepard, 1968) in combination with elevation dependent detrending. We used hourly measurements of incoming shortwave radiation from a station 12km away from the basin. Longwave radiation has not been measured in near vicinity, why it was parameterized from temperature, relative humidity and sunshine duration data measured at the same station as shortwave radiation following methods by Pirazzini (2000).

Regional climate projections

To generate model input data representative of future climate conditions we concentrated on changes in temperature, precipitation and long wave radiation. These parameters were expected to have the greatest influence on snowpack characteristics. Possible changes in relative humidity, wind speed and short wave radiation are more uncertain and were not taken into account. To estimate future climate conditions, we used RCM simulations distributed through the PRUDENCE database (Christensen et al., 2007) of daily minimum and maximum temperatures, daily accumulated precipitation and daily averages of incoming longwave radiation. These simulations were available for a control period (1961-1990) and for a future period (2071-2100) with a spatial resolution of 50km. We downloaded data for (1) the nearest grid cell, (2) for the two IPCC scenarios of greenhouse gas emission A2 and B2, and (3) simulated by the following six models: HIRAM (Danish Meteorological institute), HAD (Norwegian Meteorological Institute), PROMES (Universidad Complutense de Madrid), RCAO (Swedish Meteorological and Hydrological Institute), HDRM3 (Hadley Centre) and REGCM (Abdus Salam International Centre for Theoretical Physics of Weather and Climate Section).

Manipulation of observed data

We modified the observed data series of temperature, precipitation and longwave radiation to represent climatic conditions towards the end of the 21st century by largely following the methods described in López-Moreno et al. (In press).

Precipitation: First, the daily precipitation sums were pooled into 12 x 6 x 3 classes according to 12 months, 6 RCMs, and 3 projections (current conditions, scenario A2 and B2). Each of these 216 classes was subdivided into decile ranges (i.e. the lowest, second-lowest … highest 10% percent of data) of which we determined the mean value. These 2160 decile range mean (DRM) values were averaged over the six RCM subsets resulting in a 12 x 3 x 10 matrix. We describe the difference between the two future scenarios and the current conditions by dividing the DRM values for each scenario by the respective values representative of the current conditions. This procedure finally results in 10 DRM quotients per 12 months per 2 scenarios (Figure 1., left panel). To generate model input data for the two scenarios we manipulated measured data using these DRM quotients. Therefore, this time the measured precipitation data were pooled into monthly classes. For each of these 12 classes we calculated the decile ranges and assigned each measured values to its specific range. This allowed manipulating each measured precipitation value by multiplying it with its month-, deciles range-, and scenario-specific DRM quotient.

To modify the incoming longwave radiation measurements, we mainly applied the same method as for precipitation with two exceptions. First, we determined DRM values of daily averages instead of daily sums. Second, we described the effect of climate change with differences instead of quotients between the two future scenarios and the current conditions (Figure 1., right panel).

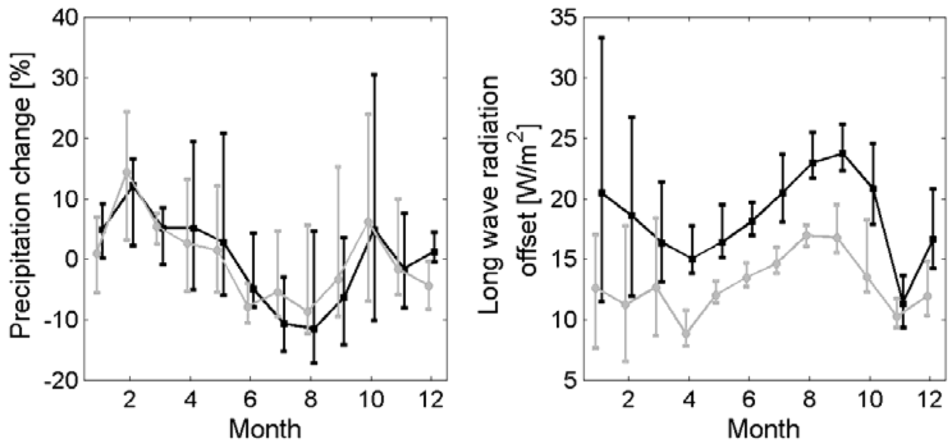

Figure 1. Change in precipitation (left panel) and longwave radiation (right panel) due to climate warming for scenario A2 (black line) and scenario B2 (grey line). The line shows the change between the control period and the two emission scenarios for the fifth decile for each month in the year, and the bars shows the corresponding change for the first and the tenth deciles.

As temperature features a considerable daily cycle, we extended the above-described methodology to additionally account for the change in daily temperature range. We therefore modified the highest and lowest daily temperatures separately (Figure 2.). Finally, the temperatures in-between the already changed daily maximum and minimum values were determined in two steps. First, for both the originally observed as well as the already changed daily extreme values, a linear change was assumed for the temperatures lying in-between. Second, the differences between these two curves were added to the observed record of hourly temperatures.

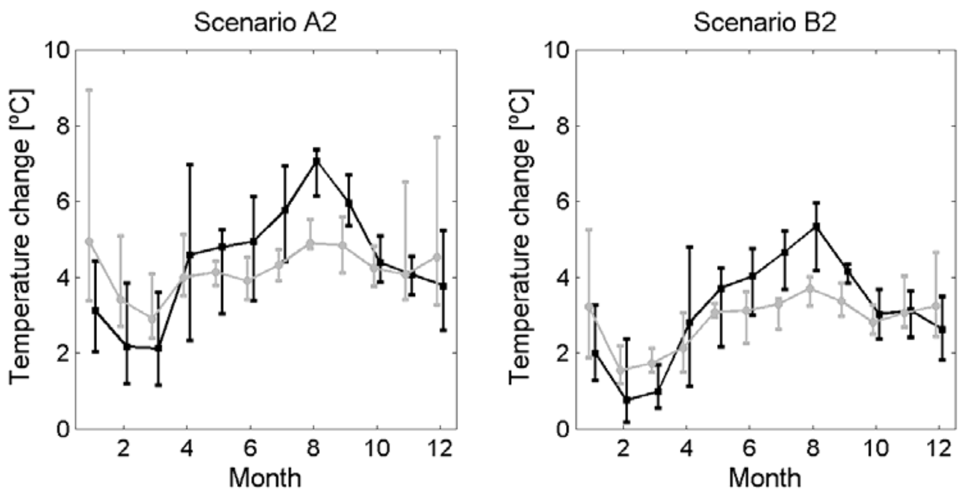

Figure 2. Change in maximum (black line) and minimum (grey line) temperatures for scenario A2 (left panel) and B2 (right panel). The line shows the change between the control period and the two emission scenarios for the fifth decile for each month in the year, and the bars shows the corresponding change for the first and the tenth deciles.

RESULTS

Validation of the present state snow simulations

We validated the modelled snow depths by comparing simulation results for current conditions against independent measurements of snow depth (Figure 3.). As no snow depths have been measured within the basin, we used data from five different stations located 5 to 15km away from the catchment, and ranging from 1440 to 2287m a.s.l. (only available after 1998). The measured snow depths were interpolated to the same grid used for the Alpine3D simulations again applying the interpolation software WINMET. First, we compared the average snow depth for all grid cells of flat pixels (slope angle less than 15°) below 2300m (Fig 3, left panel) because the interpolation is not able to reproduce snow depths in regions with complex terrain features or at altitudes outside the range of the input data. Second, we compared simulated and interpolated snow depths at a set of single grid cells representing flat field conditions for different altitudes below 2300m a.s.l. (Fig 3, right panel) to ensure that there is no systematic deviation between different elevation bands. The comparison shows good agreement between modelled and interpolated snow depths. The simulations represent dynamic changes well, but overestimates the peak snow depths for winter 1998/99 and underestimates the snow depths of the two last winters slightly.

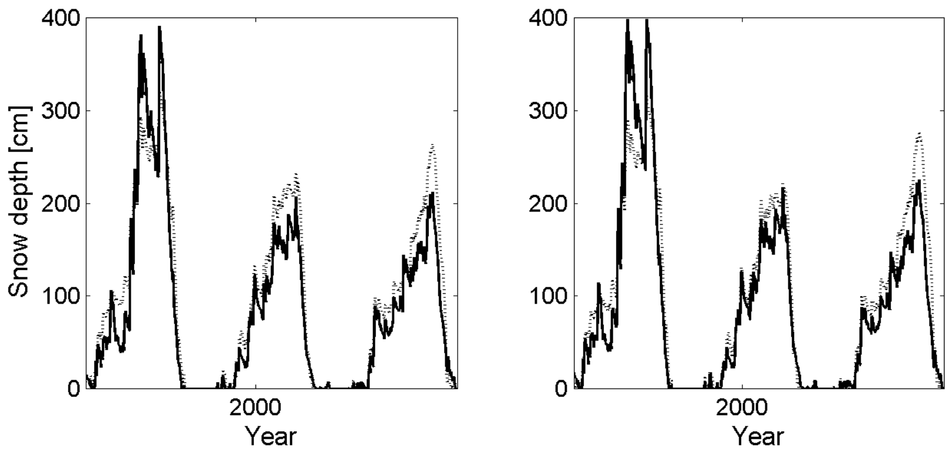

Figure 3. Modelled snow depths (filled line) and interpolated snow depths (dotted line) averaged over the flatter regions of the basin below 2300m (left panel) and for a representative flat point at 2200m altitude (right panel).

Climate change characteristics

Predicted changes in precipitation show a strong seasonal pattern (Fig 1, left panel). Winter precipitation is expected to increase by up to 14% in February. In summer, however, the climate models simulate decreased precipitation sums by around 12% in July and August. Both emission scenarios provide similar precipitation estimates. For longwave radiation, on the other hand, a raise throughout the whole year is predicted with significant differences between the scenarios (Figure 1., right panel). The offsets lie between 11 and 24 W/m^2 for scenario A2, and between 9 and 17 W/m^2 for scenario B2. They correlate to the temperature raise, which varies differently throughout the year for the daily highest and lowest temperatures (Figure 2.). In August, the increases of the maximum temperatures (7˚C for scenario A2 and 5˚C according to scenario B2) are greater than for the lowest temperatures. In winter, on the other hand, the raise of the minimum temperatures (5˚C for scenario A2 and 3˚C for scenario B2) are higher than for the maximum temperatures. Thus, the monthly spread in temperature increases for the summer, whereas it decreases during winter.

Snow covered area and mean snow water equivalent

Today, the whole basin is snow covered from mid November until mid May (Figure 4). Even in summer 14% of the catchment remains snow covered, and contributes to glacier accumulation. The peak in average snow water equivalent of 820 mm occurs during the first part of May. The accumulation period starts in the beginning of October, and the ablation period ends in late August.

For scenario A2, the catchment is completely snow covered from mid December until mid in April. Thus, the snow season begins almost one month later, and ends at least one month earlier than today. For scenario B2, the corresponding snow covered season lasts from the beginning of December until the beginning of May. In summer, the basin will be all snow free independent of the emission scenario of greenhouse gases. For scenario A2, the highest snow water equivalent occurs in the later half of April, and is 22% lower than today. The timing of the highest snow water equivalent is identical for scenario B2, and decreases 14% compared to present conditions. The accumulation period starts in the end of October, and the ablation period ends in the beginning of August for both emission scenarios. The accumulation of snow seems to remain at a similar speed, while the melting rate will become much faster.

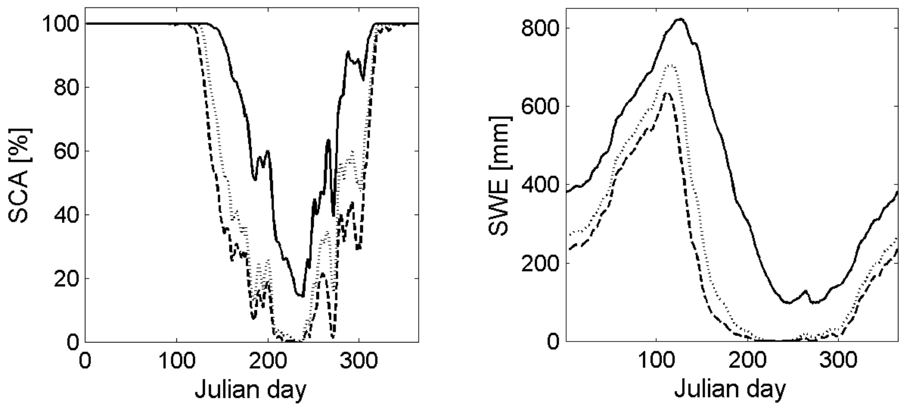

Figure 4. The effect of climate change on snow covered area (SCA, left panel) and on the mean snow water equivalent (SWE, right panel). Current conditions (filled line), scenario A2 (dashed line) and scenario B2 (dotted line).

Variation of peak snow water equivalent with altitude

The yearly peak snow water equivalents increases with altitude and occurs later in the season with higher elevations because lower temperatures at the uppermost part of the basin allow a longer snow accumulation period (Figure 5). For the present conditions, the highest snow water equivalent varies from 700 to 1950mm depending on altitude, and occurs between the end of April and the end of September. Certain areas in the elevation band from 3250 to 3400m are very steep, and therefore hold less snow (Figure 5, left panel).

For scenario A2, the highest snow water equivalents vary from 390 to 925mm depending on altitude, and occur between the end of March and the beginning of June. For the less pronounced scenario B2, the values are slightly higher, from 530 to 1030mm, and the timing is later as well, from mid April until mid of June.

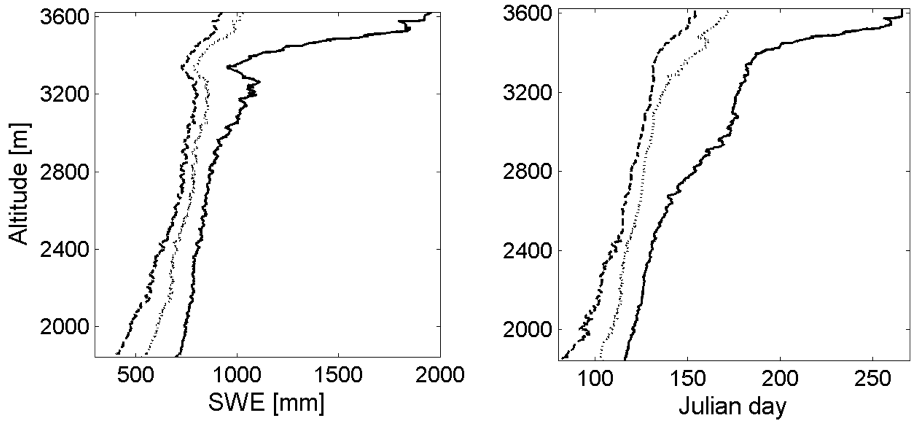

Figure 5. The effect of climate change on amount (left panel) and timing (right panel) of the maximum snow water equivalent against altitude. Current conditions (filled line), scenario A2 (dashed line) and scenario B2 (dotted line).

Duration of a seasonal snow cover for different altitudes

The duration of a seasonal snow cover is longer at higher altitudes because of the negative temperature gradient with elevation (Fig. 6). Today, the snow season starts late in September on altitudes just below 3350m, and in the beginning of November on the lowest elevations of the basin. The snow season ends in mid August for the upper parts of the basin, and in the end of May for the lowest parts. On altitudes above 3350m the basin is completely snow covered throughout the year. Some areas between 3100 and 3350m are also always covered with snow (which is not visible in the average figure). On these high elevations, the snow cover contributes to glacier accumulation.

For scenario A2, the snow season starts in late November at low elevations, and in the first part of October for the highest areas of the basin. Thus, the snow season starts between two to four weeks later than today. The snow season will also end earlier in future, between the beginning of May and the mid of July, which is between one and two months earlier than today. The changes are not as drastic for scenario B2, where the snow season starts between beginning of November and October. Further, the basin is no longer snow covered after mid May at lower altitudes, and after the end of July for the higher parts. For both future scenarios, the snow cover in the complete basin melts-out leaving no zones for glacier accumulation.

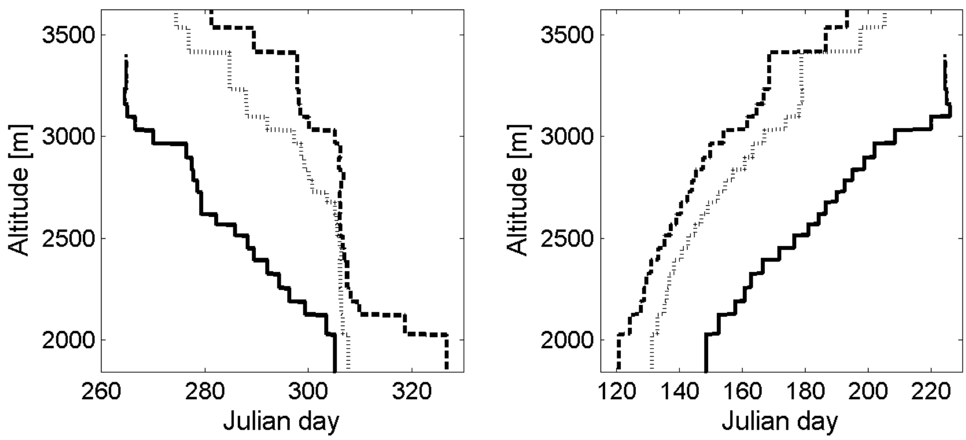

Figure 6. The change in timing from that a seasonal snow cover is built up (left panel), to its final ablation (right panel) with altitude due to climate change. Current conditions (filled line), scenario A2 (dashed line) and scenario B2 (dotted line).

DISCUSSION AND CONCLUSIONS

When carefully calibrated, conceptual snow cover models using a day-degree factor approach can efficiently and well simulate the spatial and temporal distribution of a snow cover (Zappa et al., 2003). However, such models may not be appropriate for climate change studies, as the assumption of a stationary calibration may not be valid on longer time scales, especially as global warming seems to accelerate in the 21st century (IPCC, 2007). In contrast, models describing internal snow processes that simulate the surface albedo in dependence of snow grain types seem most capable of representing the energy budget of the snowpack well (Etchevers et al., 2004). This makes them suitable for estimating trends in the snowpack response to climate change.

Alpine3D is such a model, which resolves interactions between the terrain and the radiation budget, such as the effect of albedo on the spatial distribution of long- and shortwave radiation. Also the heat fluxes at the lower boundary between the snowpack and the ground are taken into account. Therefore the effect of a glacier beneath the seasonal snow cover is modelled as well.

One of the strengths of Alpine3D is its ability to reproduce the snow cover in steep alpine terrain with the exception of effects of avalanches (Mott et al., In review), which is not the case for the interpolation of measured snow depths used as comparison. A perfect match between modelled and interpolated snow depths can therefore not be expected. Nevertheless, we find a reasonable agreement between the two estimates for flat field conditions. Discrepancies are minor and may stem from methodological shortcomings of the interpolation method.

Assuming that our modifications of the observed meteorological records are representative of future climate characteristics, the simulation results show drastic changes in both the temporal and spatial patterns of the snow cover in the Damma glacier basin. The snow season shortens with almost one month at the beginning, and with between one to two months at the end of the winter compared to the conditions of today. Similar trends in the decline of the snow season have been predicted for several other catchments in Switzerland as well (Beniston et al., 2003b, Jasper et al., 2004). Although, the decreasing number of snow covered days are not severe as on lower altitude sites, where the average winter temperatures oscillates around zero (Hantel and Hirtl-Wielke, 2007). The combined effect of higher temperatures and longer snow-free periods induces higher evaporation rates during summer, and lower accumulated sublimation during winter (Dankers and Christensen, 2005). A shift in timing, and a decrease in runoff can thus be expected. Further, in the Damma glacier basin the most marked impacts of climate warming on the snowpack will occur in the uppermost parts of the basin. Here, a raise in temperature leads to a total ablation of snow, and consequently the present accumulation area of the glacier will no longer be active in the summer. However, at these high altitudes the winter temperatures will still remain below zero, allowing a significant snow accumulation. The increase in precipitation during January to March also leads to faster snow accumulation rates compared to today. Thus, there will still be snow available during the spring melt period. Snowmelt from the region today occupied by the glacier will also contribute to runoff with a faster response time in the future. The simulations shows further that the ablation period begins a couple of weeks earlier, and that the melt rate becomes much faster. We can thus expect a quite high contribution to runoff from snowmelt, concentrated on an even shorter time period than today. On the other hand, during the summer months significantly lower runoff can be expected, as there will be no glacier or snow melt left, and also because precipitation decreases and evaporation increases. With a decreasing total snow water equivalent for all times during the year the annual runoff will decrease.

The impact of climate change also depends on the expected raise of greenhouse gases. As a simplistic plausibilization of our predictions, the elevational profiles for the maximum snow water equivalent are to be shifted upwards by approx. 500m to match the current conditions (Figure 5.). This corresponds to a raise in temperature by around 3.5˚C assuming a vertical temperature lapse rate of 0.7˚C, which is between the mean annual temperature increase expected towards the end of the 21st century for scenarios A2 and B2 (Fig. 2).

A future goal of this work is to study the effect of glacier retreat in the Damma glacier basin from a hydrological perspective. Therefore, we will extend our simulations by coupling Alpine3d to models especially designed to predict glacier mass balance changes and runoff formation in ungauged alpine catchments.

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THERMAL DYNAMICS OF SUB-ARCTIC LAKE LAGARFLJÓT, ICELAND

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ABSTRACT

Water temperature plays an important role in lake environments. Not only does temperature affect the chemical composition of water, but thermal processes also heavily influence the fate and transport of materials. This paper characterizes the thermal dynamics of Lake Lagarfljót, one of Iceland´s largest and deepest lakes. Results from 10 year temperature records suggest that despite low summer temperatures and strong winds, a weak summer stratification lasting two months sets a stage for unusually slow internal seiches. The bottom stirring and oxygen renewal from these seiches is reinforced by sediment rich plunging inflow. Differential heating and cooling may contribute to lateral transport of landborn urban materials from littoral to pelagic regions of the lake. This work highlights the importance of further study of sub-arctic lake thermal processes, which will be conducted in the coming years.

INTRODUCTION

Water temperature plays an extensive role in regulating lake environments. From a static viewpoint, water temperature affects the density, solubility of dissolved oxygen, toxicity of pollutants, and metabolism and respiration of flora and fauna and is therefore a crucial water quality parameter (Stefan et al. 1998). From a dynamic viewpoint, water temperature generates and influences the transport of materials in lakes in a myriad of ways (e.g. Imberger, 1985): First, summer stratification limits the exchange between upper and lower portions of the lake, which contributes to oxygen depletion in bottom waters in eutrophied lakes (Wetzel, 1983). Second, the interaction of wind and stratification gives rise to internal basin scale waves (called seiches) which are responsible for water and hence oxygen renewal along lake bottoms (Jónsson et al., 2002). Third, differential heat capture generates a density gradient which is known to effectively move materials from shallow littoral areas to deeper pelagic areas in lakes (Andradóttir and Nepf, 2001). Fourth, the temperature and density of inflowing river relative to lake temperature can determine the fate of land born nutrients and contaminants in lakes (Andradóttir and Nepf, 2000). For example, if the overall density of the river inflow is higher than that of the lake, it will submerge and generate water renewal in the deeper regions of the lake and limit nutrient supply in the top layers (e.g. Aðalsteinsson et al., 1992). If the river water density is lower than that of the lake, the river enters the lake surface without renewing deeper waters.

From the discussion above, it is clear that dynamical thermal processes are key determinants of physical lake behavior. Thermal stratification is arguably the one most important physical feature, as it affects eutrophication, dissolved oxygen budgets, internal wave movements and fate of river inflows. Annual stratification cycles have been excessively studied in temperate lakes which are most commonly dimictic, stratifying both in summer and winter (Wetzel 1983). Such lakes start to stratify in April and are well stratified by early June, when average temperatures in the epilimnion may exceed 20°C while hypolimnion water stays around 4-8°C. Convective cooling deepens the epilimnion in fall until a strong wind event overturns the lake (typically in November). As the lake continues to cool and ice starts to form, an inverse stratification is formed where bottom water stays around 4°C (maximum density of water) and colder (hence lighter) water resides above.

To date, thermal dynamical processes in sub-arctic aquatic environments such as Iceland, with weak summer radiation and strong winds, have received little attention. Questions remain on the response of such weakly stratified systems to wind, the relevance and presence of internal waves, density driven processes such as exchange flows driven by differential heating and cooling and inflow dynamics. This paper tackles these questions by analyzing existing temperature records from Lake Lagarfljót, one of Iceland´s deepest and longest lakes.

The paper organization is as follows: The site and monitoring program are first introduced. Then meteorological conditions and thermal processes characterized. Lastly the results and next steps are summarized.

MONITORING PROGRAM

Site description

Lake Lagarfljót is one of the largest and deepest lakes in Iceland with a surface area of 53 km^2 and a maximum depth of 112 m. Located at 65° latitude in Eastern Iceland, it classifies as a sub-arctic lake. The glacial river Jökulsá í Fljótsdal feeds the lake from the South. With the recent damming of another glacial river, Jökulsá í Dal, the two rivers now flow into the lake at same location.

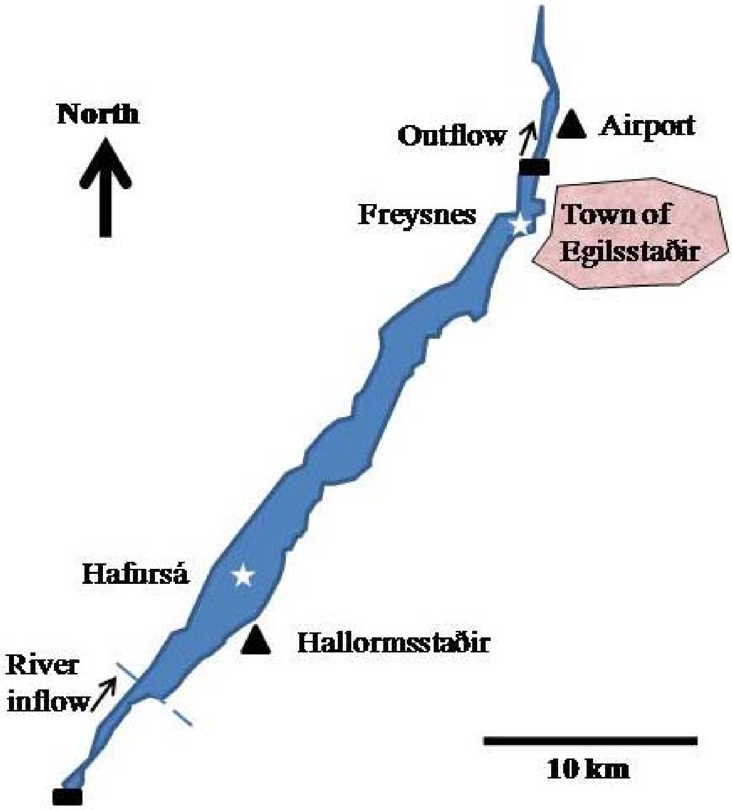

Figure 1. Lake Lagarfljót site with weather (▲), vertical water temperature (*), inflow and outflow monitoring stations (■).

Field measurements

Hourly air temperature and wind measurements are conducted by the Iceland Meteorological Office at two locations along the lake: At Egilsstaðir airport situated on the Northern end, and at Hallormsstaðir located along the Southern shore (Fig. 1).

Vertical water temperature has been monitored by the National Energy Authority of Iceland in Lake Lagarfljót since 1995 (Axelsson, 2006). Profiles are taken 10 times over the warming season with a handheld temperature probe (until 2002) and Seamon TD probe (after 2002) at two different stations: In the 100 m deep section 8 km from the southwestern end of the lake by river Hafursá, and at the 40 m deep Freysnes located close to the town of Egilsstadir in the Northern end of the lake. Hourly river inflow and outflow temperatures have been auto-recorded at 2 m depth since 1995 and 2001 respectively.

River flowrates entering and exiting the lake have been measured by the National Energy Authority of Iceland since 1962 (Jónsson et al. 1999). In addition, suspended sediment load at the inlet has been recorded since 1966 (Harðardóttir and Þorláksdóttir, 2004) and in the outflowing waters since 1995 (Harðardóttir et al. 2006).

RESULTS AND DISCUSSIONS

Meteorological data

Average monthly air temperatures typically range from -3 to -4°C in winter to 11-12°C in summer. Air temperatures change from year to year but are similar at the two measurement stations along the lake.

Wind also follows a seasonal cycle, with slightly higher wind speeds in the winter than summer. The prevalent wind direction is from the North and Northwest and secondarily from Southwest, or along the main axis of the lake. Unlike air temperature, the wind at the two measurement sites differs substantially. Fig. 2 shows that the monthly average wind speed at Egilsstadir airport is approximately 2 m/s higher than in the more inland Hallormsstadir. This spatial variability is a result of wind sheltering, as Hallormsstadir is sheltered by local mountains and to some extent by tree cover, while the airport is located in an open area away from mountains and trees. The representative wind acting over the lake likely is here assumed to lie in between the Airport and Hallormsstadir records.

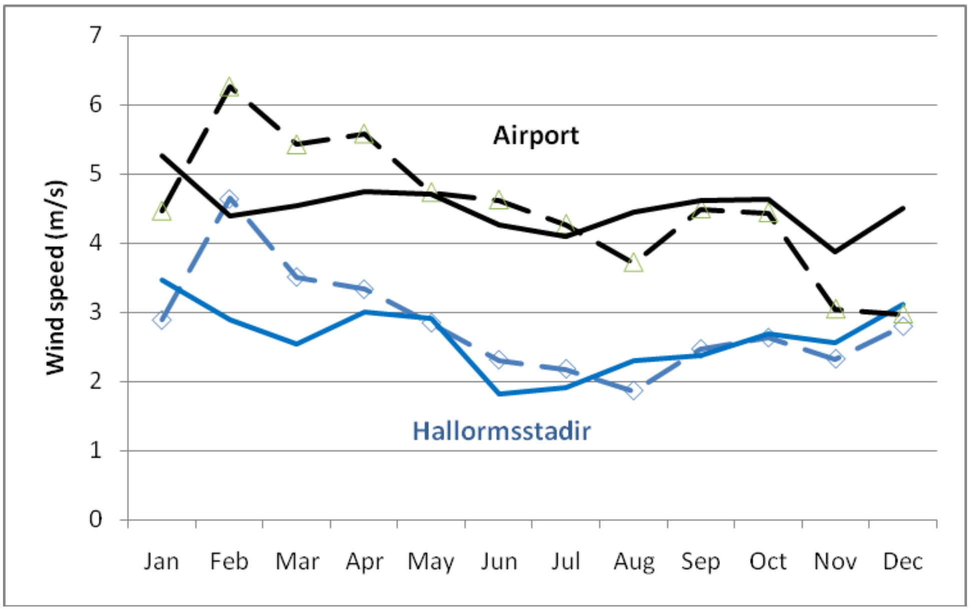

Figure 2. Wind speeds at Egilsstadir airport and Hallormsstadir in 2003 (dashed) and 2005 (solid).

Thermal stratification cycle

The annual thermal stratification cycle is arguably the principal physical characteristic feature of lakes and as such, highly important to understand. Figure 3 summarizes the time variation in vertical temperature in the deep portion of Lake Lagarfljót at Hafursá from April to November 2005. The solid lines depict isotherm of constant temperature in °C constructed from individual profiles taken nine times during the season (see crosses on x axis). In April and May, the lake water temperature is uniform over depth just above freezing point. By June, the water has warmed and reached the temperature of maximum density (4°C) and the lake is inversily stratified, where 4°C warm water resides at the bottom of the lake under colder lighter water. This inverse stratification appears to be shortlived, because in July the water temperatures exceed 5°C and the surface is slightly warmer than bottom layer. By the end of July, the lake is weakly stratified with bottom water hovering around 6°C and surface water exceeding 8°C. The lake remains stratified until the middle of September. In mid October, the lake has overturned and is uniformily 6°C warm. To summarize, the 2005 measurements suggest that the sub-arctic Lake Lagarfljót is inversily stratified for some days in June and then continues to become weakly stratified in August which lasts through mid September.

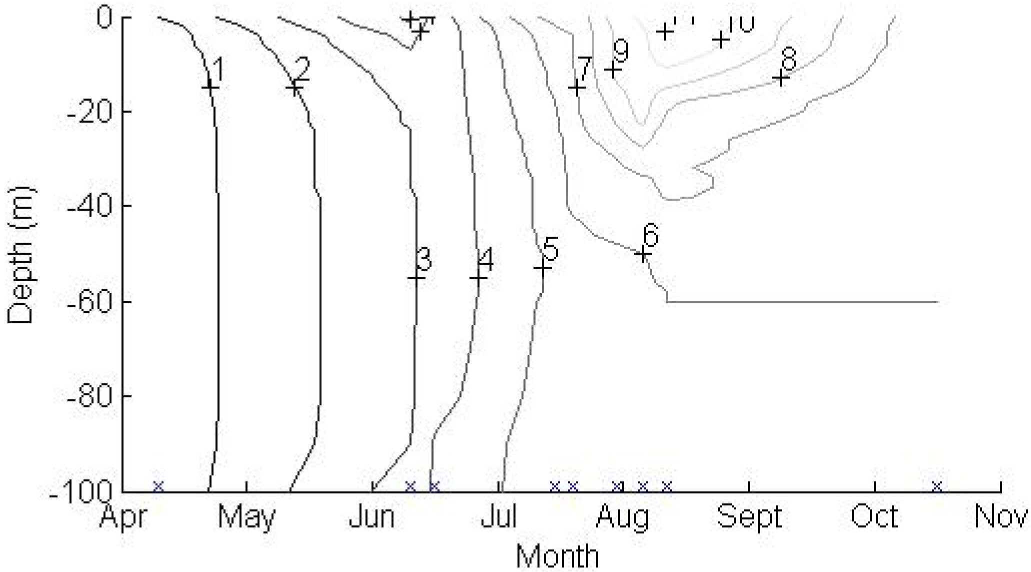

Figure 3. Depth-time diagram of isotherms (°C) in Lagarfljót 2005

Water temperature measurements from other years suggests that the summer stratification may develop as early as the end of June and last through the end of September. The longest observed period of summer stratification is hence 3 months, but an average timeframe is conservatively estimated as 2 months. Vertical temperature differences range from 2-5°C, the strongest stratification occurring consistently in August.

This annual temperature evolution is in sharp contrast to the well studied temperate lakes in North America and Europe described in the introduction of this paper: First, the summer stratification is very weak, with a vertical temperature gradient on the order of 0.1°C/m, which is ten times smaller than in many temperate lake systems (Wetzel, 1983, p. 75). Second, the summer stratification lasts only for 2 months as opposed to 7 months. Third, the onset and destruction of stratication occurs very quickly as opposed to a slow buildup in spring and thermocline deepening over several fall months, consistent with the weak stratification. Last, inverse stratification occurs in

June just prior to summer stratification, and it is shortlived. While few temperature records exist during peak winter, it appears that the sub-arctic lake does not stay inversily stratified for a long period of time because winter temperatures drop uniformily below 1°C. This differs from dimictic lakes which can be inversily stratified for months in the wintertime.

One of the key water quality concerns associated with stratification is the suppression of vertical mixing between upper and lower water layers, ε_z. In strongly stratified eutrophic lakes, the lack of mixing contributes to oxygen depletion of lower waters. Research have shown that vertical mixing is inversily related to the buoyancy frequency, N (Fisher et al, 1979, p. 195), i.e.

$$\varepsilon_z \propto (N^2)^{-a} = \left(-\frac{g}{\rho}\frac{\partial \rho}{\partial z} \right)^{-a}$$

Since the vertical temperature gradient is small and only lasts for two months, then vertical mixing is not as severly suppressed in sub-arctic Lake Lagarfljót as in temperate lake systems. This is supported by the temperature measurements on Fig. 3, which shows that bottom temperatures change substantially over the course of the year. Still, this weak stratification may generate internal wave activity and impact the fate of river inflow. This is investigated in the next sections of the paper.

Wind driven seiches

Theoretically, wind generates both surface and internal seiches in stratified lakes. Wind strength is characterized by its shear stress, τ_w, which is related to the wind speed measured at 10 m height, W_{10}^{2}, and the direction between main wind speed, θ, and axis of the lake, θ_x, i.e.

$$\tau_w \approx \rho 10^{-6} W_{10}{}^{2} cox(\theta - \theta_x)$$

The amplitudes of wind generated seiches can be derived analytically as (Gill, 1982, p. 122)

$$\eta_s = \frac{\tau_w}{\rho g H_s}\frac{L}{2} \qquad \eta_i = -\frac{\rho}{\Delta\rho}\eta_s$$

where H_s is the depth of surface layer, L the lake length, and $\Delta\rho$ the density difference between the epilimnion and hypolimnion waters. The first order periods of these seiches can be determined as (Wetzel 1983)

$$T_s = \frac{2L}{\sqrt{gH}} \qquad T_i = \frac{2L}{\sqrt{\Delta\rho/\rho \, g \, H_s H_b / H}}$$

Using the observed temperature difference between 15 m deep epilimnion and hypolimnion of 2-5°C, monthly average wind speeds, lake length of 25 km, surface seiches are expected to have amplitudes ranging in mm and period of 40 min. The internal seiches, however, may have amplitude of 5-10 m depending upon the stratification and wind. Their periods of oscillation are estimated to range between 3-6 days, which is very slow compared to internal periods in temperate lakes which are typically on the order of hours (Wetzel, 1983). Future research aims to measure these slow moving wind driven internal seiches and improve the understanding of movements along the bed which are important to benthic flora.

Differential heating and cooling

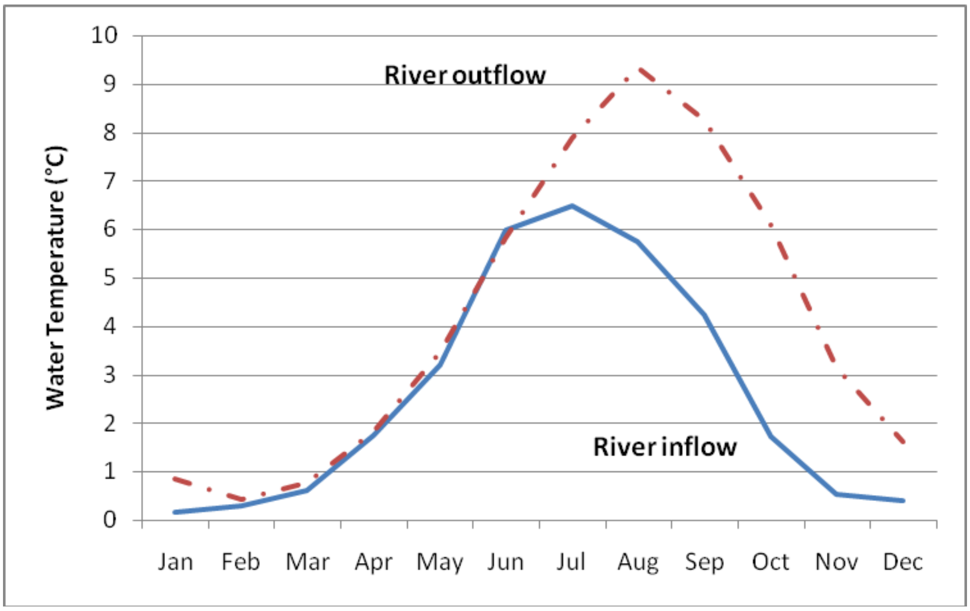

Figure 4. Average monthly water temperatures of inflowing river (Hóll) and outflowing river (Lagarfell) 2001-2007.

Figure 4 shows the monthly average water temperatures in the river both entering and exiting the lake. The water temperatures mimic each other closely until July, when the exiting water becomes 1-4°C warmer than the river entering the lake. This suggests that the lake water, and in particular, the water in the shallower region at the northern end of the lake where the outflow occurs, is warmer than the glacial river inflow. When comparing two vertical temperature profiles inside the lake (Fig. 5), it becomes clear that differential heating and cooling occurs where the 40 m deep Freysnes warms (or cool) more than the 100 m deep Hafursá. Considering more profiles, a temperature difference of 1-2°C is typical within the top 40 m. A lake bathymetry map shows that the basin deepens steadily from 40 m to 90 m over a 9 km distance which can be taken as the representative length scale (L) for the lateral density gradient $\partial\rho/\partial x \approx \Delta\rho/L$.

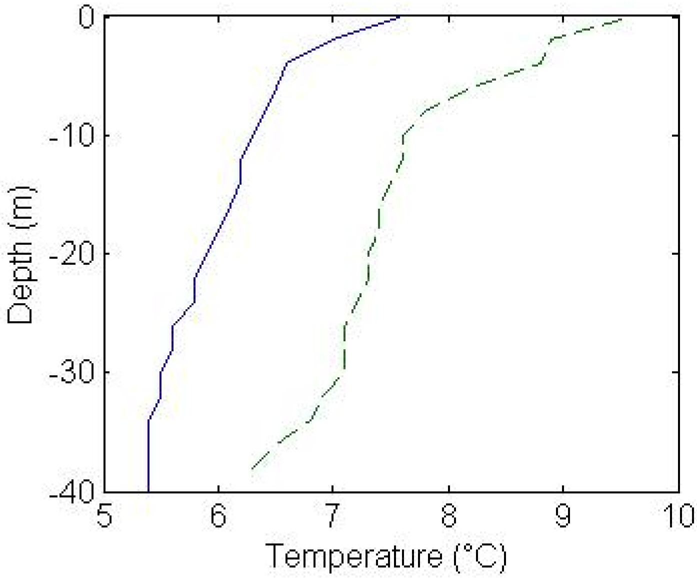

Figure 5. Vertical temperature profile at Hafursá (solid) and Freysnes (dashed) July 14, 2005 5-7 PM.

This gradient is large enough to generate exchange flows that drive landborn materials from shallow to deep regions in the lake, which can be seen by comparing the strength of density gradient ($\partial\rho/\partial x$) relative to the wind (τ_w) through the longitudinal gradient Wedderburn number (Andradóttir and Nepf, 2001)

$$We_g = \frac{g \, {\partial\rho}/{\partial x} \, H^2}{\tau_w}$$

Here H is the effective depth and g is gravity acceleration. If |We|>>1, then differential heating and cooling is the dominant important lateral transport mechanism, while if |We|<<1 then wind is the dominant lateral transport mechanism.

Using the observed density gradient over the top 40 m and average monthly summer wind speeds (2-4.5 m/s) yields a Wedderburn number in the range of 5-40. This suggests that thermally driven exchange flows may be an equally (if not more) important transport mechanism for moving materials from shallow to deeper sections of the lake as wind. In the absense of wind, lateral transport is still present because of these currents. Hence any landborn materials coming for example from the town of Egilsstadir may reside for a long time in the lake, because the density gradient moves them towards the interior of the lake as opposed to the nearby river outlet.

Inflow dynamics

The fate of river inflow in lakes depends on the density of the river water relative to the lake water. Since temperature affects water density, the temperature of river relative to distribution of temperature of the lake impacts the river intrustion depth. Fig. 6.a) shows the monthly average river temperatures relative to the range of temperatures typically observed at different depths inside the lake. The river appears to be slightly warmer than the lake during spring and somewhere in between epilimnion and hypolimnion temperatures during summer. As the lake water stores more heat, the glacial river becomes and remains colder than the lake throughout the end of the year. In the absense of suspended sediments, this would indicate that the river water enters the surface or medium depths from March through July.

For glacial rivers carrying large sediment loads, water density is also a function of suspended sediment concentration. Tómasson´s et al. (1996) sediment rating relationship suggests that the suspended sediment concentration of the river inflow to Lake Lagarfljót, C_s (in g/l or kg/m^3), can be estimated from historical river flowrates, Q (in m^3/s) (Jónsson et al. 1999), as

$$C_s = 0.007Q^{0,95}$$

The results on Fig. 6.b) indicate a strong seasonality in suspended sediment concentrations in the inflowing glacial river. During the glacial melting season from May through September, suspended sediments exceed 200 mg/l which is approximately tenfold the concentration measured in the lake outflow, which stays around 20-40 mg/l range throughout the year (Tómasson´s et al. 1996; Harðardóttir et al. 2006). Using the stirred reactor concept, one can stipulate that the lake outflow reflects lake concentrations. Fig. 6.b) hence suggests that lake inflow concentration is greater than the lake outflow most of the year.

The combined effect of temperature (ΔT) and suspended sediment concentrations (ΔC_s) on density anomaly is calculated as

$$\rho_R - \rho_L = -\alpha\Delta T + \left(1 - \rho_w / \rho_p\right)\Delta C_s$$

Here α is the thermal expansivity of water, ρ_p the weight of inorganic sediments (2700 kg/m^3) and ρ_w the weight of water (1000 kg/m^3) following Finger et al. (2006). The result shown on Fig. 6.c) indicates that in summer, the high suspended sediment concentration in inflowing river overpowers any temperature effects resulting in that the heavier river plunges into the lake even if it may be warmer. Over the peak winter months, however, Fig. 6.c) indicates that the river water may be lighter than the lake and hence enter at the surface.

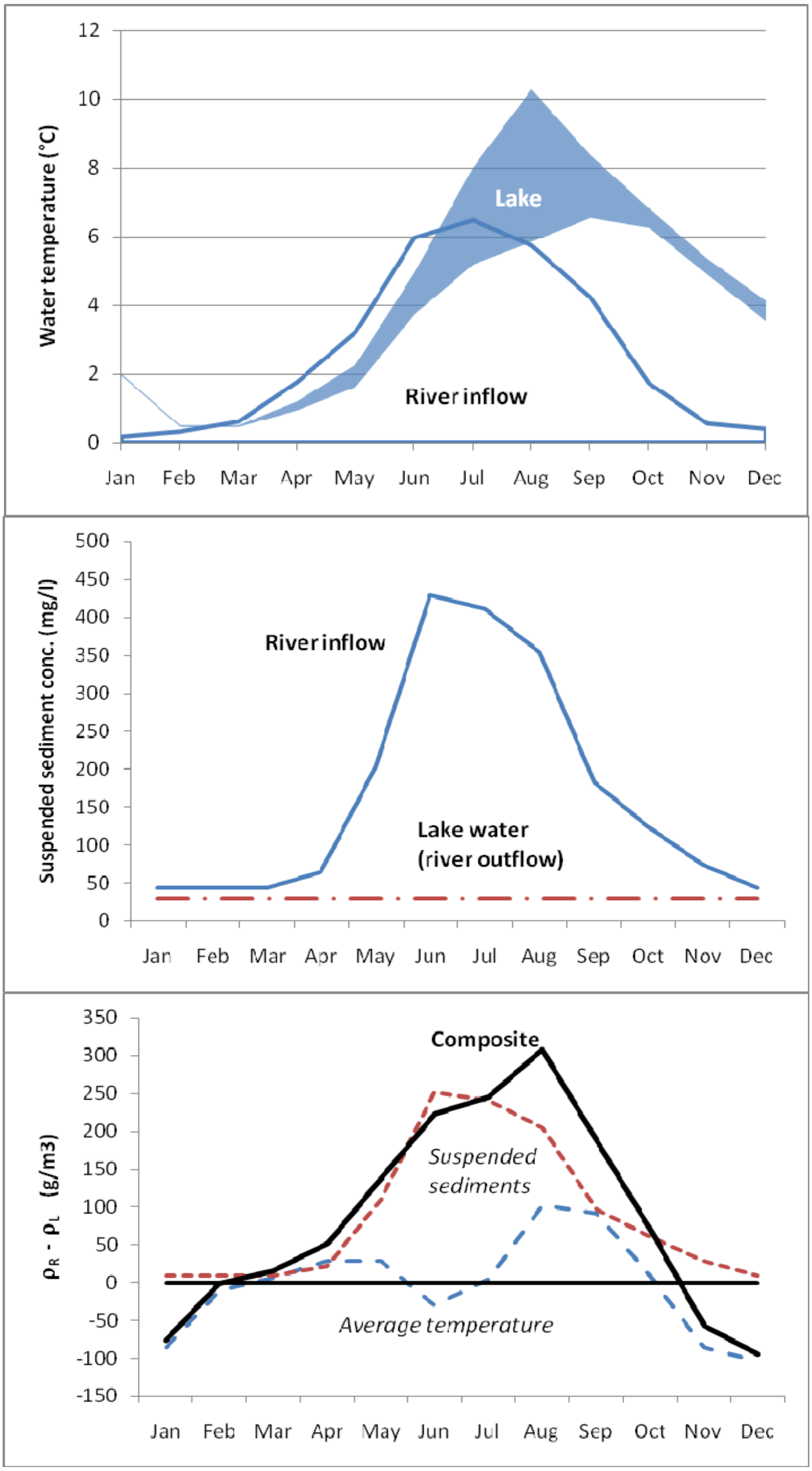

Figure 6. Monthly average a) water temperature (2001-2007) and b) suspended sediment concentration of river inflow relative to lake (1963-1997), and c) calculated density difference.

This analysis suggests that lake inflow dynamics may follow a strong seasonal cycle, with plunging river inflows in the summer and surface intrusions in winter. This has important implications on the water renewal of bottom waters, as the river inflow brings dissolved oxygen to the deeps of the lake in summer while little renewal occurs in the winter, which may lead to oxygen depletion.

SUMMARY

This paper demonstrates that a sub-arctic lake exhibits interesting density and wind driven behavior despite low summer temperatures and limited solar radiation. Ten years of temperature measurements in Lake Lagarfljót show that the lake weakly stratifies during the summer for 1.5-3 months. While the stratification is weak (only 0.1°C/m) and does not suppress vertical mixing to the same extent as temperate lakes, it may set a stage for slow moving internal seiche motions with periods on the order of several days, which is considerably longer than for better studied temperate lakes. These wind driven basin scale waves may be important to the livelihood of benthic flora and fauna. The fate of river inflow follows a seasonal cycle. In summer, sediment rich glacial river inflow plunges into the lake, reinforcing water renewal along the lake bottom boundary, which is often the critical concern in lake water quality. In the wintertime, the colder yet lighter river water intrudes the lake surface. This change in inflow dynamics suggest that oxygen depletion is most likely to occur in wintertime, when the lake is frozen (and hence not moving with wind) and inflow is entering at surface. Lastly, differential heating and cooling is prevalent in this sub-arctic lake and may contribute to lateral transport from littoral to pelagic regions, in combination with wind. The work presented in this paper is an important stepping stone for future research, which will focus on better characterizing the wind generated internal waves and near bed bottom movements in weakly stratified sub-arctic systems.

ACKNOWLEDGEMENTS

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THE INITIATION AND DEVELOPMENT OF A JÖKULHLAUP FROM THE SUBGLACIAL LAKE BENEATH THE WESTERN SKAFTÁ CAULDRON IN THE VATNAJÖKULL ICE CAP, ICELAND

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ABSTRACT

Results from investigations of a jökulhlaup from the subglacial lake beneath the Western Skaftá cauldron in the Vatnajökull ice cap are reported. Following thermal drilling into the lake in June 2006, the lake temperature was measured and the elevation of the ice shelf covering the lake recorded with a permanent GPS station. A number of parameters were monitored in the Skaftá river where jökulhlaups from the cauldron regularly emerge. Subsidence of the ice shelf was recorded by the GPS instruments during a jökulhlaup from the Western cauldron in September 2006. We estimate that 64 Gl of water emptied from the cauldron during this jökulhlaup, which reached a maximum discharge of 150 m^3/s. The maximum discharge at the glacier snout was 135 m^3/s and the speed of the flood front under the ice cap was found to be in the range 0.2–0.4 m/s.

INTRODUCTION

Skaftárkatlar are two circular depressions, 1–2 km in diameter and up to 150 m in depth, located in the northwestern part of Vatnajökull. They are formed by steady subglacial melting due to the presence of powerful geothermal areas beneath each cauldron (Björnsson, 2002). The melting sustains 100 m deep subglacial lakes beneath 300 m thick ice cover (Jóhannesson et al., 2007). Jökulhlaups regularly flow into the river Skaftá when the meltwater escapes from the cauldrons. The period between jökulhlaups from each cauldron is 2–3 years and about 40 events are known during the last half century (Zóphóníasson, 2002). The total volume discharged in a single jökulhlaup averages 0.1 km^3 from the western cauldron and 0.25 km^3 from the eastern cauldron.

These jökulhlaups are of the rapidly rising type, normally reaching maximum discharge in 1–3 days and then receding in 1–2 weeks (Björnsson, 2002). This type of jökulhlaup behaviour stands in contrast to the typical, slowly rising outburst floods from the Grímsvötn subglacial lake, which have been explained by the classic Nye theory of jökulhlaups (Nye, 1976). The Nye theory fails to simulate the rapidly rising jökulhlaups from Skaftárkatlar (Björnsson, 1992).

An extensive campaign involving measurements within the cauldrons, in the subglacial lakes and of the jökulhlaups originating in them, was initiated in 2006. The 300 m thick ice shelf covering the western cauldron was penetrated by a hot water drill in June 2006 (Thorsteinsson et al., 2007; Jóhannesson et al., 2007). Temperature profiles in the lake were measured and a water sample taken for geochemical and microbiological studies. A pressure and temperature sensor was deployed at the bottom of the lake and connected with a cable to a continuously recording datalogger at the surface. A continuously recording differential GPS instrument was placed at the center of the ice shelf and a water temperature logger was placed in the Skaftá river 3 km from the port where the river emerges from beneath the ice. In addition, data from the Sveinstindur hydrological station in Skaftá, located 25 km from the glacier margin, have been used in this study. Below we outline results collected during a jökulhlaup from the western cauldron in late September 2006.

METHODS AND DATA

The discharge of Skaftá at Sveinstindur can be calculated from a known relationship between the discharge and the water level recorded by the hydrological station. Water temperature, conductivity, light absorption and the air temperature are also recorded. The jökulhlaup is well resolved in this data set, which shows a rising water level starting from the night before September 27th (Fig. 1). The conductivity meter was unfortunately not working during the jökulhlaup but normally conductivity is higher for jökulhlaup water then for normal glacial meltwater, because of the added geothermal component. Because of substantial warming of jökulhlaup waters during flow from the glacier snout to the Sveinstindur station, located 25 km downstream, we cannot rely on data from the station as a source of information on the temperature of the water emerging from the subglacial water course. Thus it was decided to place a temperature logger in the river course 3 km from the glacier snout, selecting a location where the logger would not be washed away with the jökulhlaup. The logger became submerged in the evening of September 28th (Fig. 2). It was placed above the normal winter surface of the river and thus did not become submerged until the water level had risen considerably.

Figure 1. Discharge measured at Sveinstindur during the jökulhlaup. Labels on the x-axis denote the beginning of the corresponding day.

Figure 2. Water temperature measured in Skaftá 3 km from the glacier snout. Measurements before the sharp drop on September 28th are air temperature measurements, as the temperature sensor did not become submerged in water until on the evening of the 28th of September.

The distance from the glacier snout to the gauge at Sveinstindur affects the timing and the shape of the flood wave at the gauge location. This causes problems in estimating the speed of the flood front under the glacier, total storage of water under the glacier and other characteristics of the subglacial flood wave. To overcome this problem the one-dimensional hydraulic model HEC-RAS was used to calculate the expected shape and timing of the jökulhlaup at the glacier snout (Fig. 3) (Jónsson, 2007). In addition to the jökulhlaup component, the discharge measured at Sveinstindur includes the normal glacial discharge component in Skaftá and tributary rivers. Here the Skaftá discharge is assumed to be constant at 55 m^3/s and tributary rivers entering the western branch of Skaftá between the glacier snout and Sveinstindur are assumed to contribute 20 m^3/s.

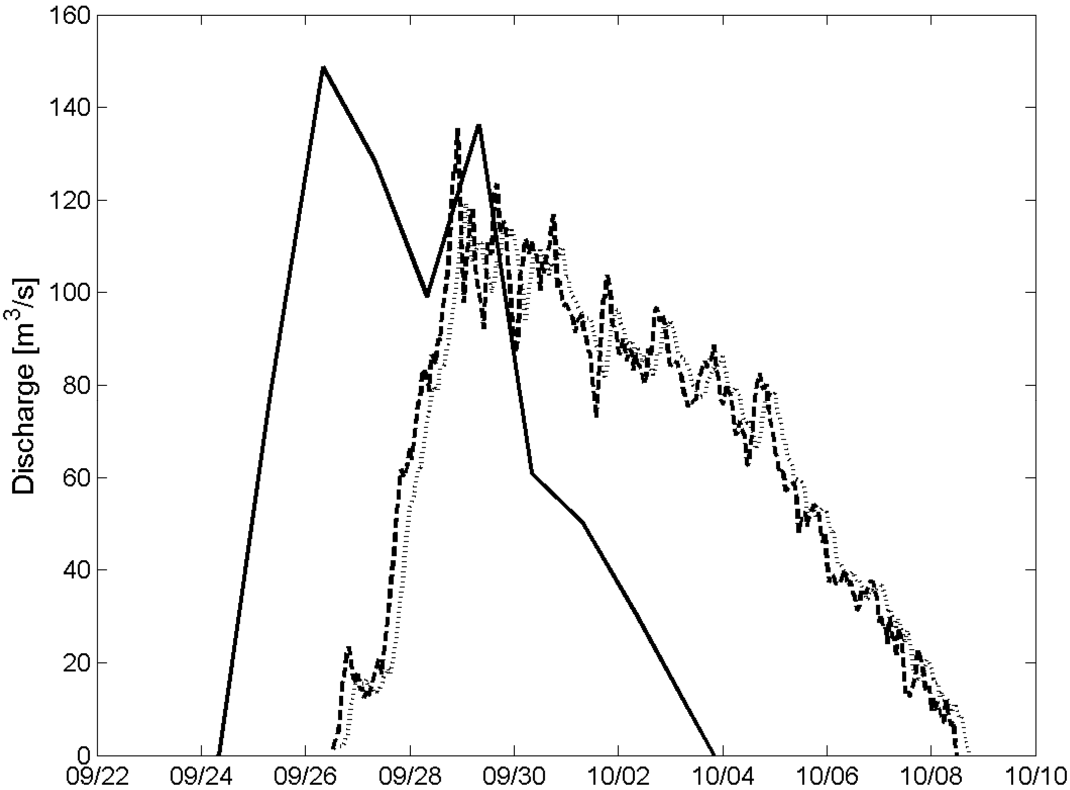

Figure 3. Discharge of flood water during the jökulhlaup. Discharge out of the cauldron is shown as a solid line, discharge at the glacier snout is shown as a broken line and discharge at Sveinstindur is shown as a dotted line.

To monitor the water accumulation in the subglacial lake and outflow during the jökulhlaup a GPS instrument that records and stores position and elevation once per day was placed at the centre of the ice shelf covering the western cauldron. Data on the water level increase was also obtained from a pressure transducer deployed at the bottom of the lake, which sent data via cable to a recording station at the surface. The cable broke in mid-September due to increasing strain in the ice shelf and thus no pressure data are available from the transducer during the jökulhlaup. Comparison of the two data sets indicated a good match between results from the GPS recorder and the pressure transducer until the cable broke and thus we may expect the GPS data to provide accurate information after that.

A relationship between the volume of the subglacial lake and the elevation of the overlying ice shelf was derived from information about the shape of the cauldron when the lake is empty. This made it possible to convert the lowering of the ice shelf to outflow of flood water from the lake. The glacier bottom is expected to be reasonably smooth and the lake assumes the form of a half dome extending upwards from the glacier bed into the ice. Prior to a jökulhlaup, the water is kept sealed in the lake by a minimum in the water potential due to the surface depression formed by melting at the base (Björnsson, 2002). The cauldron will thus take the shape of an inverted lake when the lake is emptied.

RESULTS

The 2006 jökulhlaup was comparatively small for the western cauldron with a maximum discharge of 120 m^3/s of the outburst water at Sveinstindur, 135 m^3/s at the glacier snout and 150 m^3/s outflow from the cauldron (Fig. 3). The total volume of the outburst water from the cauldron was $64 \cdot 10^6$ m^3 (Fig. 4). The maximum subglacial water storage during the flood was estimated to be $38 \cdot 10^6$ m^3 by comparing the cumulative outflow from the cauldron with the cumulative outflow at the glacier snout (Fig. 4). Comparison of the estimated variation of the volume of the subglacial water storage (Fig. 4) with the flood discharge at the glacier snout (Fig. 3) shows that both reached a maximum at about the same time, indicating as expected that the subglacial pathway is able to carry the largest discharge when its average cross sectional area is at maximum.

Comparison of the timing of the flood front at the glacier snout and the initiation of subsidence within the cauldron gives the travel time of the flood front under the ice cap. As the GPS recorder in the cauldron only records elevations once per day, the timing of the start of the subsidence has an uncertainty of 24 hours. The start of the subsidence is between 8:00 on the 24^{th} and 8:00 on the 25^{th} of September. Then there is also uncertainty in the exact timing of the start of the jökulhlaup at the glacier snout, mainly because the daily discharge fluctuation masks the slow discharge increase in the beginning. The jökulhlaup can be assumed to start at the glacier snout between 13:00 and 22:00 on September 26^{th} and thus the travel time of the subglacial flood wave from the cauldron to the snout was between 29 to 60 hours. The travel distance of the jökulhlaup, measured on a watercourse map drawn by Magnússon (2003), is 39 km and thus the mean travel speed of the front of the flood wave between the cauldron and the snout was in the range 0.2–0.4 m/s.

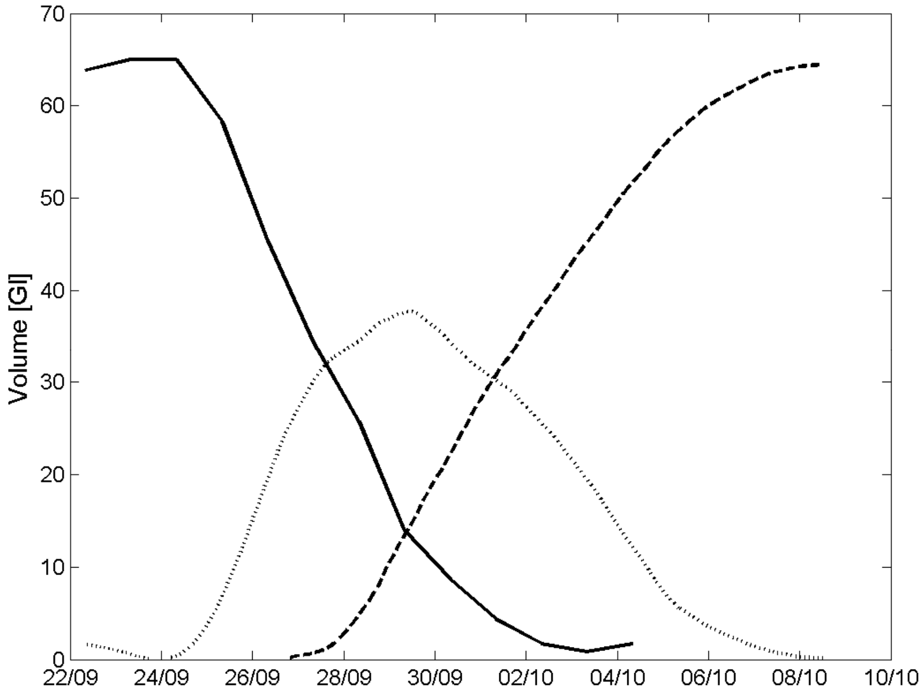

Figure 4. Volume of floodwater. Water volume in the subglacial lake shown as a solid line, cumulative volume at the glacier snout shown as a broken line and volume stored subglacially is shown as a dotted line.

At the location near the glacier snout, the temperature of the jökulhlaup water was between 0.0°C and 0.5°C (Fig. 2). The air temperature recorded at Sveinstindur varied between −1°C and 10°C during the jökulhlaup, but was between 2°C and 6°C most of the time. Hence, some warming of the outburst water may be assumed to take place on the 3 km long distance between glacier snout and the measuring point. In the morning of October 1st and during the nights of October 6th and 7th water temperatures are close to zero and the air temperature at Sveinstindur is also close to zero at the same time, so warming is probably minimal. Taken together, these measurement results indicate that the outburst water is at – or very close to – the freezing point as it emerges from beneath the ice cap. The temperature of the water in the cauldron was measured over a three month period before the jökulhlaup and found to be near 4°C (Jóhannesson et al., 2007) indicating that most of the thermal energy in the lake water was used for melting of ice on the way down the subglacial water course.

DISCUSSION

A jökulhlaup from the Western Skaftá cauldron in late September and early October 2006 was monitored with simultaneous measurements within the cauldron and in the Skaftá river carrying the jökulhlaup. These measurements make it possible to describe the variation of outflow of flood water from the cauldron and the discharge of the flood at the glacier snout and thereby time-dependent variations in the storage of water in the subglacial water course. Measurements of the temperature of floodwater near the glacier snout, furthermore, show that the water emerges from the glacier near the freezing point. In this paper, we have presented some of the data along with a preliminary interpretation, but hopefully further modelling efforts using the data will contribute to an increased understanding of the mechanisms of fast rising jökulhlaups.

The travel speed of the flood front under the glacier (0.2–0.4 m/s) is slow compared to the travel speed of subglacial water under normal conditions and flood waves in open channels. There are indications that the travel speed of the water increases at later stages for jökulhlaups in Skaftá, as exemplified by data collected after discharge had peaked during the 2002 jökulhlaup from the Eastern Skaftá cauldron. For this estimation we used data on: i) earthquake tremors indicating boiling or a minor volcanic eruption at the cauldron floor because of the pressure release accompanying the jökulhlaup, and ii) the concentration of suspended material in the jökulhlaup waters collected at the gauging station, which display a peak believed to result from the same boiling/eruption event. In this case, the flow speed under the glacier was estimated to be approximately 0.8 m/s.

Outburst water is expected to be at or very close to the freezing point as it emerges from the glacier as thermal transport from jökulhlaup water to the surrounding ice walls in the tunnel is expected to be effective (Jóhannesson, 2002). The measured data provide support for this interpretation, but future studies should aim for water temperature measurements at the point of emergence at the glacier snout, to avoid the effects of warmer air on the river temperature.

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HYDROLOGIC MODELING OF TWO CANADIAN WATERSHEDS USING THE NORTH AMERICAN REGIONAL REANALYSIS DATA

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ABSTRACT

In part due to concerns about the impact of climate change, there has been an increased interest in hydrological modeling of watersheds in Canada. Most of Canada is sparsely populated and a recurrent problem in hydrologic modeling is the lack of reliable weather data at the sites of interest. This paper explores the use of a recently released reanalysis data set for hydrologic modeling. The North American Regional Reanalysis provides temperature and precipitation on a 32 km by 32 km grid which is appropriate for hydrologic modeling. These data are used as input to the hydrological model SLURP in lieu of weather station data. For the particular cases considered here, it is found that model calibration using NARR data is quite acceptable and comparable to what can be obtained using interpolated weather station data.

INTRODUCTION

Canada possesses a large proportion of the world's freshwater stock. The management of this resource in the context of possible climate change is a significant challenge, not least because of the scarcity of streamflow and weather data in the remote regions of the country. Climate change impact assessment of water resources usually involves the use of a hydrologic model. However, the lack of data makes it difficult to develop reliable models. Often information must be interpolated from weather stations located far outside the basin of interest, a procedure which may result in poor model performance.

As an alternative to weather station data, one could consider data from reanalyses. Reanalysis data are obtained by running numerical weather prediction models that assimilate observations of variables such as surface pressure, relative humidity, precipitation, and temperature observed at different locations. The output of the reanalysis includes information about surface precipitation and temperature on a regular grid. These variables are the main input to most hydrologic models.

The objective of the present study is to compare calibrations of the hydrologic model SLURP with observed weather data and with data from the recently released North American Regional Reanalysis (NARR) data set. Two watersheds in the Winnipeg River basin in northern Ontario are modeled.

NORTH AMERICAN REGIONAL REANALYSIS (NARR)

The North American Regional Reanalysis (NARR) is a reprocessing of historical meteorological observations using NCEP's regional Eta model and various data assimilation systems. NARR's spatial domain covers United States, Canada, and Mexico, and data from 1979 – updated in near real-time – are available at high temporal (3 hours) and spatial (32 km × 32 km) resolution. One key feature of this data set is the improved assimilation of observed precipitation (Mesinger et al., 2006) which makes it superior to previous global reanalysis data sets from NCEP-NCAR (Kalnay et al., 1996) and NCEP–Department of Energy (Kanamitsu et al., 2002). While NARR climatologies have been studied in some detail, the data set has rarely been used at daily time scales for hydrological modeling. Woo and Thorne (2006) applied SLURP to the Liard basin in the mountainous region of western Canada using climate data from various sources including weather stations, NCEP-NCAR Reanalysis, and NARR. They obtained similar model performance at the basin outlet but summer peaks were poorly simulated. Although their work provides valuable information regarding the utility of the NARR data set, it is limited to a specific region and was based on a short period of analysis.

STUDY AREA

Two catchments were selected for this study: the Sturgeon River and the Troutlake River, located in the greater Winnipeg River basin in north-western Ontario, see Figure 1 and Table 1. Troutlake River and Sturgeon River gauging stations are located within 50 km from the Redlake Airport weather station and the Sioux Lookout Airport weather station, respectively. The basins are unregulated, but the runoff regime is influenced by many small lakes.

Table 1. Streamflow gauges in the study area.

Name	Period of Record	Drainage Area
Sturgeon River at McDougall Mill	1961-2005	4450 km^2
Troutlake River above Big Fall	1970-2005	2370 km^2

THE SLURP MODEL

The SLURP ("Semi-distributed Land Use-based Runoff Processes") model (Kite, 1995) is a basin model which simulates runoff based on daily weather input (precipitation, mean temperature, relative humidity, and bright sunshine hours) and physiographic data (land cover and elevation). It is a semi-distributed conceptual hydrological model that was initially developed for modeling meso-scale Canadian watersheds as an alternative to the use of larger and more complicated hydrological models (Kite, 1995). This model is known to be robust and suitable for the northern environment at various geographical scales (Leenders and Woo, 2002). The model can be used to examine the effects

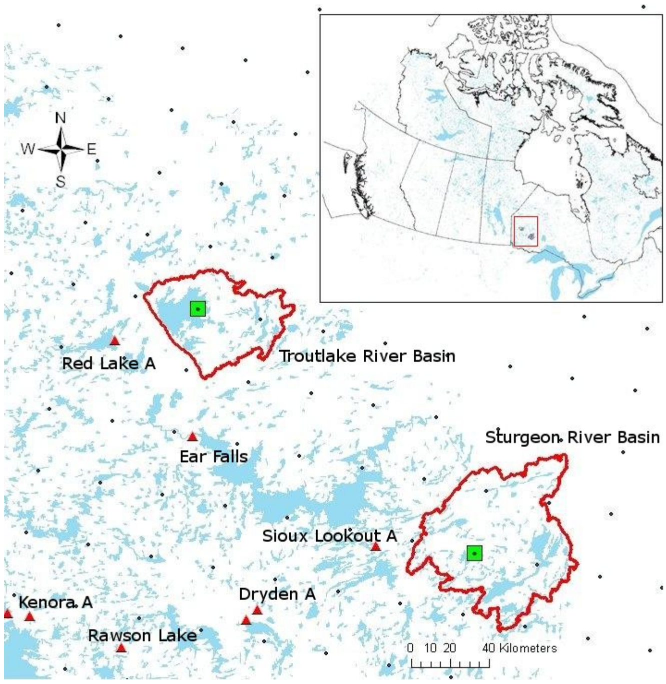

Figure 1. Location of river basins, weather stations, and NARR grid points (squared points are selected NARR grid points used for model input).

that external factors such as climate change or changing land cover might have on the hydrologic cycle.

The model divides a basin into a number of "aggregated simulation areas" (ASAs). An ASA contains certain types of land cover and the vertical water balance is calculated for each land cover in each ASA. The water is routed to the outlet of each ASA and then to the outlet of the basin. SLURP simulates the vertical water balance with four storage tanks in each land cover in each ASA: canopy store, snow store, fast store, and slow store. Precipitation is provided as input of water to ASAs, and fluxes such as interception, sublimation, evapotranspiration (ET), surface runoff, interflow, and base flow are calculated from the storage tanks.

COMPARISON OF OBSERVED WEATHER DATA AND NARR DATA

Data from the two weather stations (Redlake A, Sioux Lookout A) were used to calculate monthly climatologies of precipitation and temperature. These were compared to the corresponding NARR climatologies at the nearest grid points to provide an initial assessment of the suitability of NARR data for hydrological modeling. NARR mean precipitation from early summer through fall is lower than the observed, while NARR mean temperature is higher (Figure 2). With a few exceptions, the biases are relatively small, and one should keep in mind that weather stations may be influenced by localized effects.

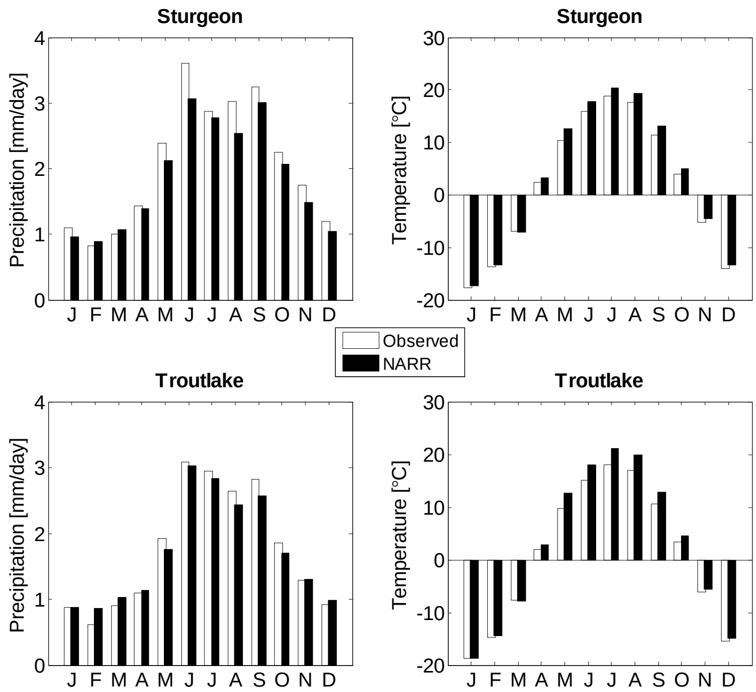

Figure 2. Mean monthly precipitation and temperature from weather stations and NARR.

RESULTS

SLURP calibration and validation using observed data

The SLURP model was set up for each basin using digital land cover and elevation data. The digital elevation data were obtained from the NASA Shuttle Radar Topography Mission (SRTM), and land cover data sets were derived from the Advanced Very High Resolution Radiometer (AVHRR). Daily meteorological variables measured at the weather stations were used as input to each model.

The models were calibrated using streamflow data measured at the outlet of each watershed. Both weather stations have missing data, so calibration and validation were conducted using the most complete periods of record. The Sturgeon-model was calibrated for the period 1992-1995 and validated over the period 2000-2004. The Troutlake-model was calibrated for the period 1994-1997 and validated over 2000-2004. The key parameters adjusted during the calibration included maximum infiltration rate, retention constant for fast store, maximum capacity of fast store, retention constant for slow store, maximum capacity of slow store, rain/snow division temperature, canopy capacity, albedo, snowmelt rate, and evaporation-related parameters such as wilting point and field capacity. The calibration criteria included deviation of volume (D_v) of mean runoff and the Nash-Sutcliff Efficiency (E) of the daily runoff series.

Table 2 presents a summary of model performance statistics for the validation period. The results from the Sturgeon-model and the Troutlake-model are reasonable both in terms of volumetric error and goodness-of-fit.

Table 2. SLURP model performance using observed data for each basin (Validation).

	Sturgeon	Troutlake
Observed mean runoff (m^3/s)	46.13	20.68
Simulated mean runoff (m^3/s)	49.14	19.86
D_v of mean runoff (%)	6.51	-3.96
E of daily runoff series	0.77	0.65

SLURP calibration and validation using NARR data

Since the NARR data do not contain any missing data, the calibration and validation periods using NARR input data can be chosen arbitrarily. However, for the purpose of comparison with observed input data, the validation period was chosen to coincide with the same period considered in the previous section. The NARR-based models were calibrated using the years 1989-2000 and validated over 2000-2004. As shown in Table 3, the simulation statistics using NARR data are quite similar to the results obtained with observed data (Table 2).

Table 3. The SLURP model performance using NARR data for each basin (Validation).

	Sturgeon	Troutlake
Observed mean runoff (m^3/s)	46.14	20.68
Simulated mean runoff (m^3/s)	45.04	19.86
D_v of mean runoff (%)	-2.38	-3.99
E of daily runoff series	0.64	0.61

Daily streamflow simulation

The quality of NARR simulated streamflow can be compared at different time scales. Climate change studies typically focus on longer time scales such as monthly or annual, with the possible exception of the study of extreme events such as floods. In this study, we considered daily, monthly, and annual flows.

Daily streamflow were simulated for each basin using NARR input for the period 1981-2004. The SLURP models calibrated with NARR were used. The simulated flows for the two watersheds are shown in Figure 3. There is a reasonable agreement between simulated and observed flows, although many

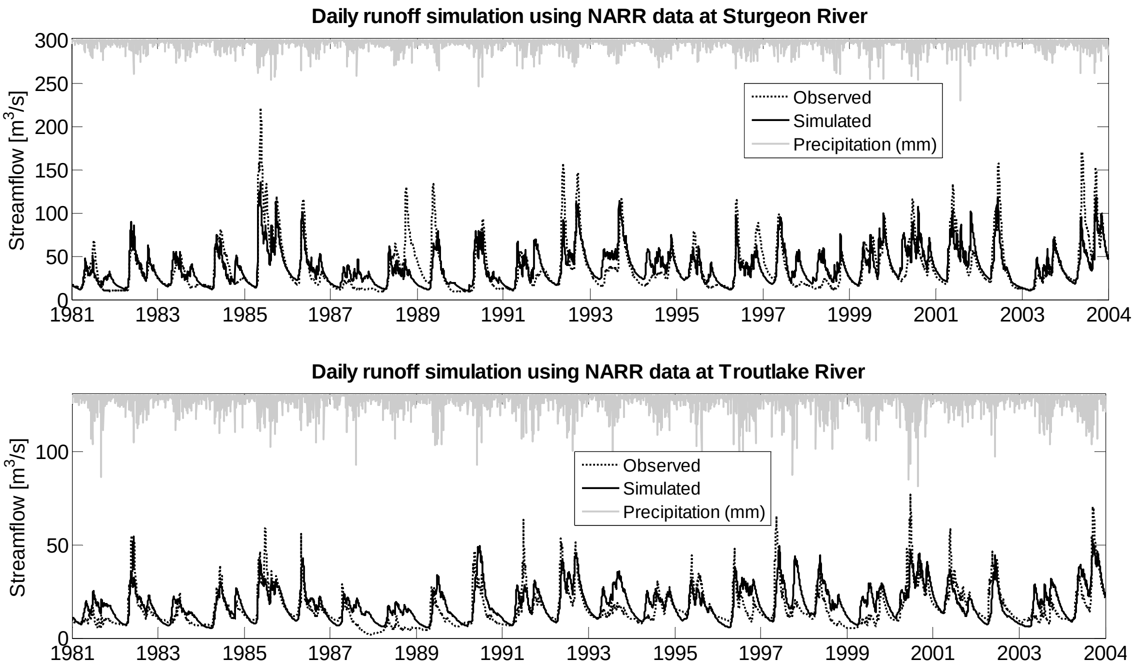

Figure 3. Daily streamflow simulation. Top: Sturgeon River. Bottom: Troutlake River.

peak flows, especially in the spring, are underestimated in the simulations. One can speculate whether the inability to adequately simulate peak flows is related to the hydrologic model or the NARR input data or a combination of the two. The climatologies shown in Figure 2 suggest that NARR has a negative precipitation bias at the Sturgeon basin (Red Lake Airport) in the month of June where much of the spring runoff takes place. On the other hand, peaks appear to be underestimated also in the case where flows are simulated with observed weather information (figures not shown). It seems reasonable to assume that the inability to reproduce many of the observed peak flows is related more to the hydrologic model than to the input data.

Monthly runoff

Observed and simulated daily flows were aggregated to monthly flows and examined in more detail. The mean monthly flows provide some insight into seasonal biases. Figure 4 shows a plot of the means of recorded monthly flows, flows simulated with observed data, and flows simulated with NARR data. For Sturgeon River, there is relatively little difference between the two set of simulated flows. Both sets of simulations underestimate mean monthly flows in May and June, a consequence of the underestimated spring peak flows seen in

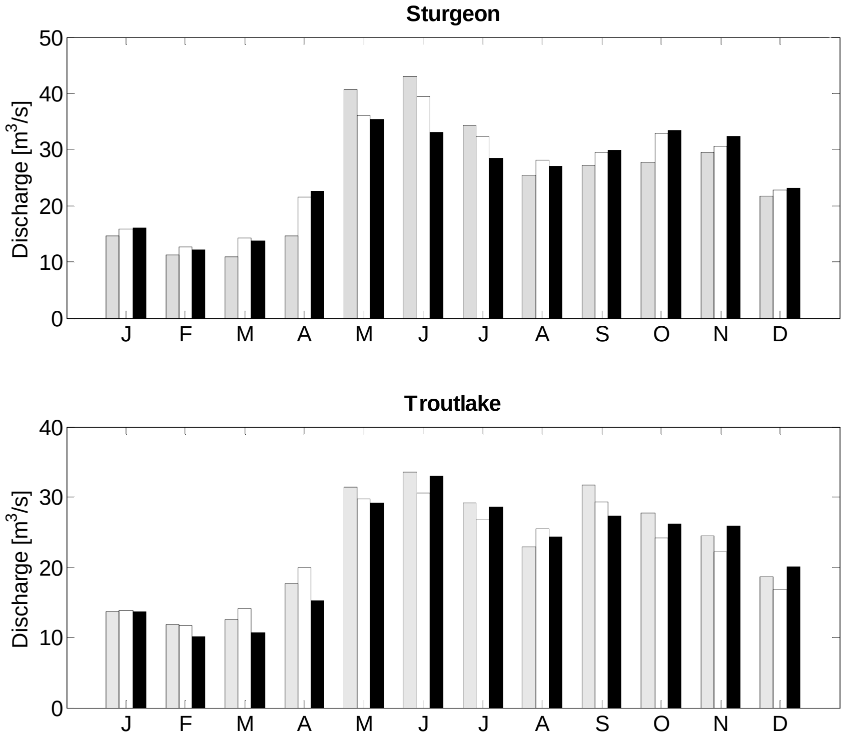

Figure 4. Mean monthly streamflow record and simulated runoff with different input data sets (observed data and NARR data).

the daily flows. The negative biases in May and June combined with the calibration criterion of minimizing the overall volumetric error leads to some overestimation in other months.

For Troutlake River, the performance of the two simulation sets in terms of monthly means is also quite similar. The underestimation of flows in May and June are somewhat smaller than for Sturgeon River.

Figure 5 shows time series of observed and simulated monthly mean runoff. Some of the problems with the simulated daily streamflows are also present in the monthly streamflows. In particular, there appears to be a tendency that months with high runoff are underestimated and months with low runoff are overestimated. Further fine-tuning of model parameters could possible reduce this problem.

Annual discharge

In the final comparison, daily flows are aggregated to annual flows to assess if the inter-annual variability of flows can be properly simulated using NARR data. The results are shown is Figure 6. For the years where reasonably complete observed weather records are available, the corresponding simulated flows are also shown. The NARR simulated flows appear to be more inaccurate than flows simulated with observed weather data. While the general variation is

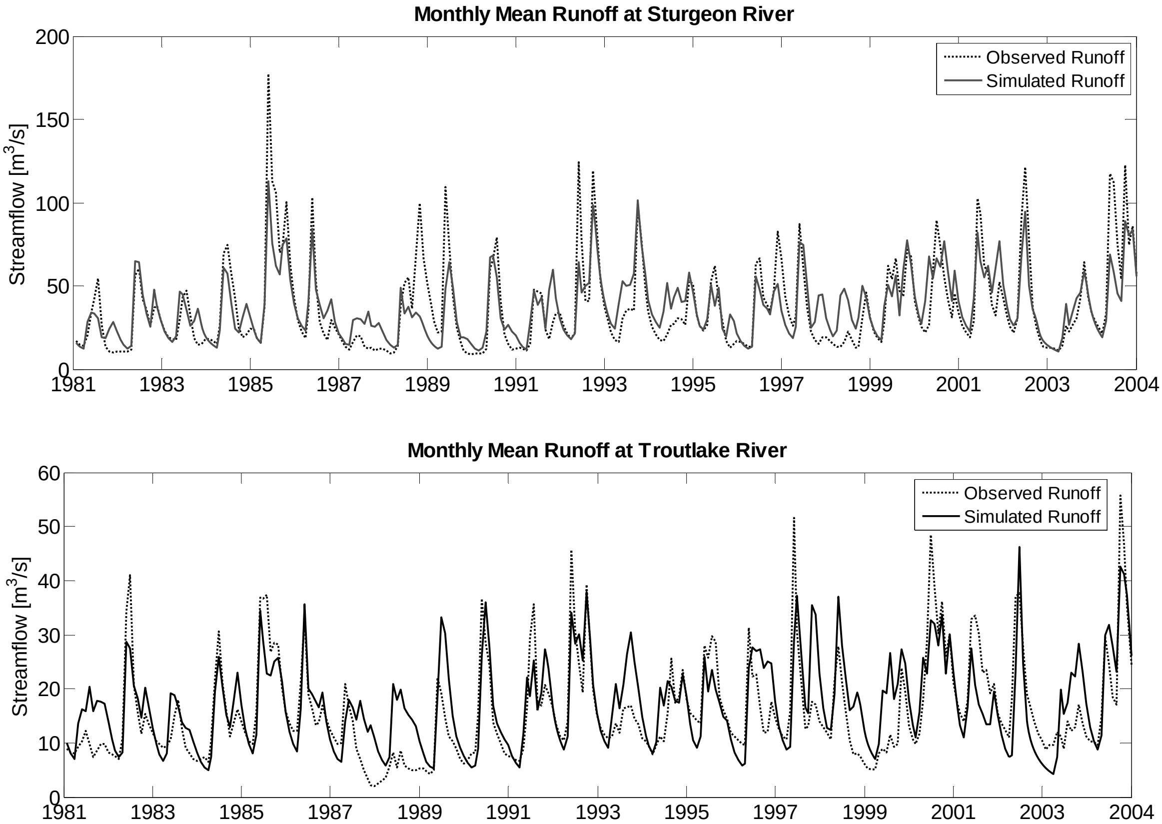

Figure 5. Observed and simulated monthly runoff series. Top: Sturgeon River. Bottom: Troutlake River.

reasonably preserved by the NARR simulation, some years are clearly not well represented.

Table 4 summarizes goodness-of-fit statistics for NARR simulated flows in the two basins, considering the different levels of aggregation.

Table 4. Model performance for different levels of aggregation. Period is 1981 to 2004.

		Daily	Monthly	Annual	Mean Monthly
Sturgeon River	E	0.62	0.66	0.60	0.89
	D_v	0.78	0.82	0.80	0.78
Troutlake River	E	0.48	0.51	0.39	0.57
	D_v	16.0	15.9	16.0	16.0

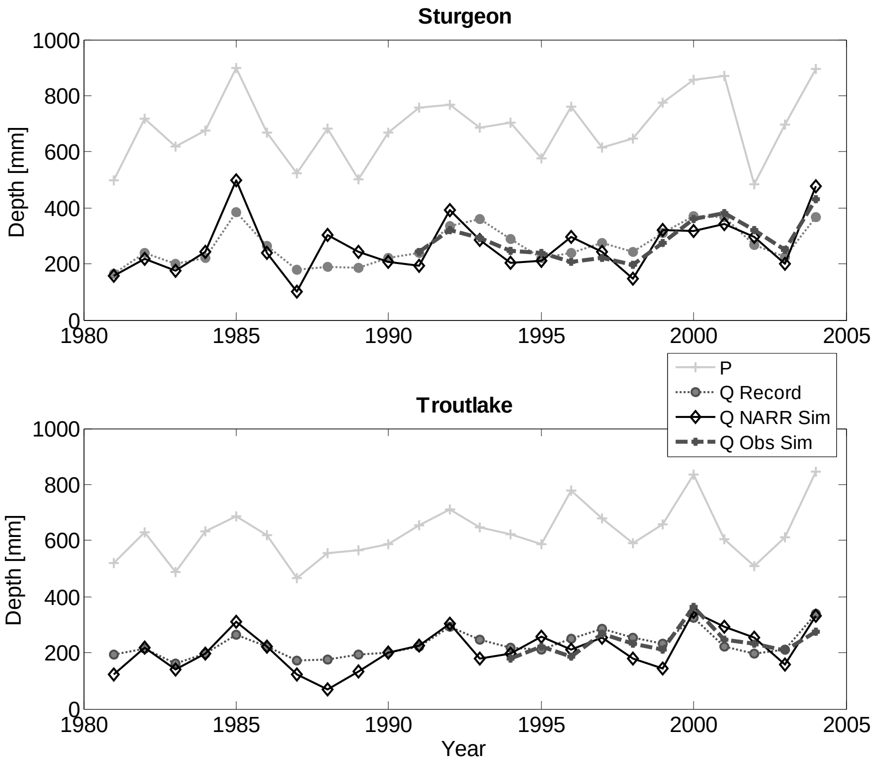

Figure 6. Annual runoff simulation for Sturgeon and Troutlake.

CONCLUSIONS

Good quality weather data are always desirable in the calibration of and simulation with hydrologic models. When representative weather data are available from weather stations in the watershed, these will in most cases be preferable. However, when precipitation information must be interpolated from weather stations located far from the basin, as is typically the case in data sparse regions, there is likelihood that weather input variables will be misrepresented. For such cases, the use of high-resolution reanalysis data may be a viable alternative. Our investigation of two watersheds in the Winnipeg River basin suggests that in a number of aspects streamflow simulated with weather information from the North American Regional Reanalysis are comparable to streamflow simulated with observed weather data. Daily and monthly data were found quite acceptable, but the representation of interannual variability was found somewhat deficient. This is an ongoing research project and future efforts will include applications to other sites as it is difficult to draw general conclusions from only two applications.

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RADIATION BUDGET IN A SNOW-COVERED SUBALPINE FOREST AND ITS INTERACTIONS WITH THE SNOWPACK

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ABSTRACT

Boreal and subalpine forests cover large areas of the Northern Hemisphere land surface. Seasonal snow is strongly influence by the presence of a forest canopy. Changes in the radiation budget and of turbulent transport processes modify accumulation and ablation of the snowpack and subsequent runoff processes. In this study we concentrate on interactions between the snowpack and the radiation budget inside the forest.

Short-wave and long-wave radiation inside forests are highly variable in space and time due to the heterogeneity of the canopy. In winter the interception of snow in trees and a non-uniform accumulation of snow on the ground lead to an even more complex radiation dynamics. Therefore, representative measurements of net radiation, albedo and transmissivity inside a snow-covered forest require an elaborate set-up.

In this study we present observational data of the radiation budget inside a coniferous subalpine forest covering four winter seasons using a novel measuring set-up. A four-component net-radiometer was mounted on a carriage and constantly moved back and forth on a 10-m rail. The device thus captured the natural variability of the net-radiation and of the forest floor albedo, respectively. Further radiation measurements above and outside the forest as well as complementary snow measurements (depth and water equivalent) enabled analysing the reciprocal effects of the radiation budget on the snowpack and vice versa.

ICE AND WEATHER CONDITIONS IN THE GULF OF RIGA, THEIR IMPACT ON RIGA HARBOR WINTER NAVIGATION, 1996-2006

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ABSTRACT

Sea ice is very important hydrological element of the Baltic Sea environment, directly or indirectly affecting many of the oceanographic, climatic, ecological, economical and human parameters that characterize the region. Winter navigation requires an icebreaker's assistance to provide safe and effective navigation in ice in the Gulf of Riga. Analysis of maximum ice extent in the Baltic Sea within 1901-2006 period shows continued tendency to decrease, however several winters with quite severe ice conditions for navigation were observed in period 1996-2006.

INTRODUCTION

The Baltic Sea is partly covered with ice every winter, mainly the Gulfs of Bothnia, Finland and Riga. The maximum annual ice extent is between January and March, when ice covers 52,000-420,000 km² and on average 218,000 km² (Lindqvist and Gullne, 2006). The Gulf of Riga freezes almost every 4 years (Pastors et all, 1996).

In winter period in the Gulf of Riga, fast and drift ice can be observed. Fast ice exists in coastal areas but drift ice has dynamic nature being forced by winds and currents. Ridges and brash ice are most significant obstructions to navigation (Pastors et all, 1996).

Winter navigation requires icebreaker (for the Gulf of Riga – "Varma") assistance to provide safe and effective navigation in ice. It is very important also for gulf's ports work.

ICE CONDITIONS IN THE GULF OF RIGA

Ice formation in the Gulf of Riga starts in the Bay of Pärnu where new ice forms in mid-December (Fig. 1).

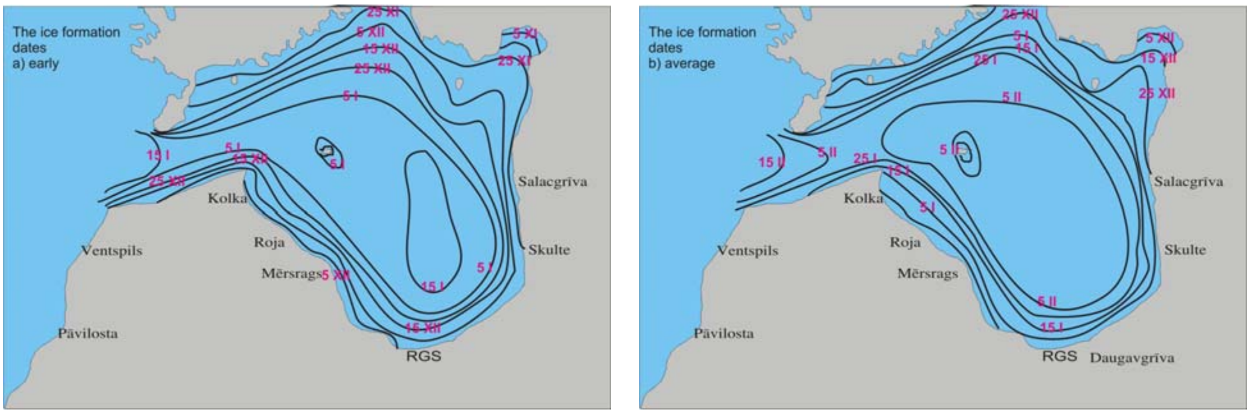

Figure 1. The ice formation dates (I. Mikelsone), a) early; b) average.

In February, intensive ice formation and freezing can be observed in the whole gulf. In severe winters, the gulf is entirely covered with ice in January. In very mild winters, the gulf does not freeze at all (Pastors et all, 1996).

The formation of fast ice usually begins at the coast, and further extends in parallel with isobaths. Maximum development of the fast ice falls on the end of February – beginning of March (Wang et all, 2003).

Some pictures of the ice cover in the gulf were taken in ice observing flight on 30 March, 2005 (Fig. 2 and Fig. 3).

Figure 2. Central part of the Gulf of Riga (E. Lépy).

Figure 3. Irbe straight (E. Lépy).

Ice processes can be described according to 7 distinguished zones of freezing probability (Fig. 4).

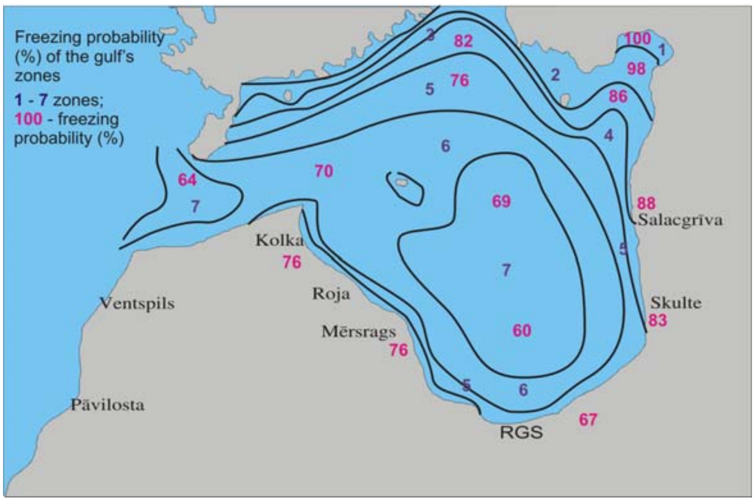

Figure 4. Freezing probability (%), of the gulf's zones (I. Mikelsone).

Winter classification in Latvia was created regarding the sum of negative daily mean air temperatures: severe, average and mild. This sum characterizes the severity of ice season in the Gulf of Riga: 330 °C corresponds to average winters; up to 670 °C the winters are severe; and below to 330 °C, it is mild (Kostjukov, 2002). This classification can show some regional differences in Latvia - in the period 1996-2006 (data were obtained from Latvian Environment, geology and meteorology agency), station Rīga (southern part of the gulf) has known 5 average and 5 mild winters, station Kolka (western part of the gulf) 2 average and 8 mild winters, whereas station Ainaži (eastern part of the gulf) has known 1 severe, 5 average and 4 mild winters (Tab. 1).

Table 1. Sum of negative temperatures, °C.

YEAR	STATION NAME		
	RĪGA	KOLKA	AINAŽI
1996/1997	-341	-180	-339
1997/1998	-255	-141	-287
1998/1999	-499	-270	-487
1999/2000	-183	-85	-183
2000/2001	-224	-166	-269
2001/2002	-271	-157	-313
2002/2003	-584	-458	-707
2003/2004	-307	-215	-351
2004/2005	-391	-153	-480
2005/2006	-582	-407	-605

Different approach for classification of winters was developing in Finland – 5 winter classes according to the maximum ice extent in the Baltic Sea (Tab. 2).

Table 2. The Baltic Sea winters classification (Leppäranta and Omstedt, 1999).

Winter class	Maximum ice extent, thousand km²	Maximum ice extent, %
Very mild	to 81	to 19
Mild	from 84 to 139	from 20 to 33
Average	from 148 to 272	from 35 to 65
Severe	from 280 to 382	from 67 to 91
Very severe	from 390 to 420	from 93 to 100

In the last decade, in the Baltic Sea region regarding to the maximum of ice extent, has been observed 5 average (Fig. 5) and 5 mild winters (Linqvist et all, 2006). This classification shows close connection with previous classification method for the Gulf of Riga.

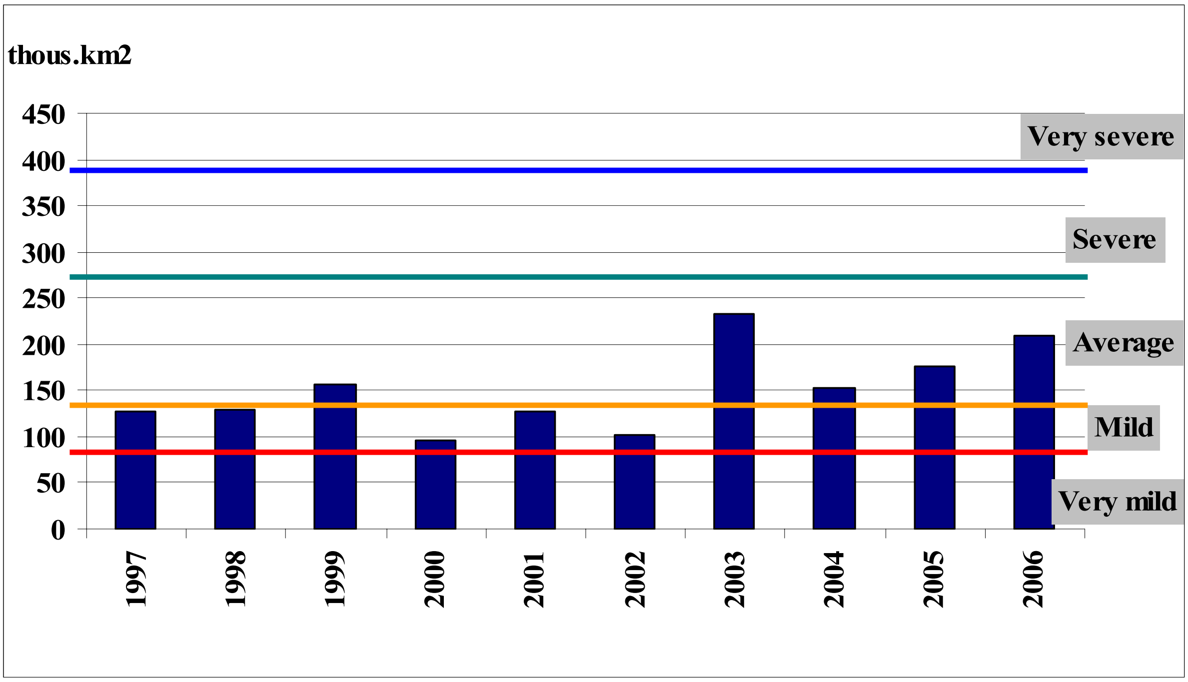

Figure 5. Maximum ice extent and winter types in the Baltic Sea, 1996-2006.

Analysis of maximum ice extent within 1901-2006 period shows continued tendency to decrease (Fig.6), however several winters with quite severe ice conditions for navigation were observed in 1996-2006.

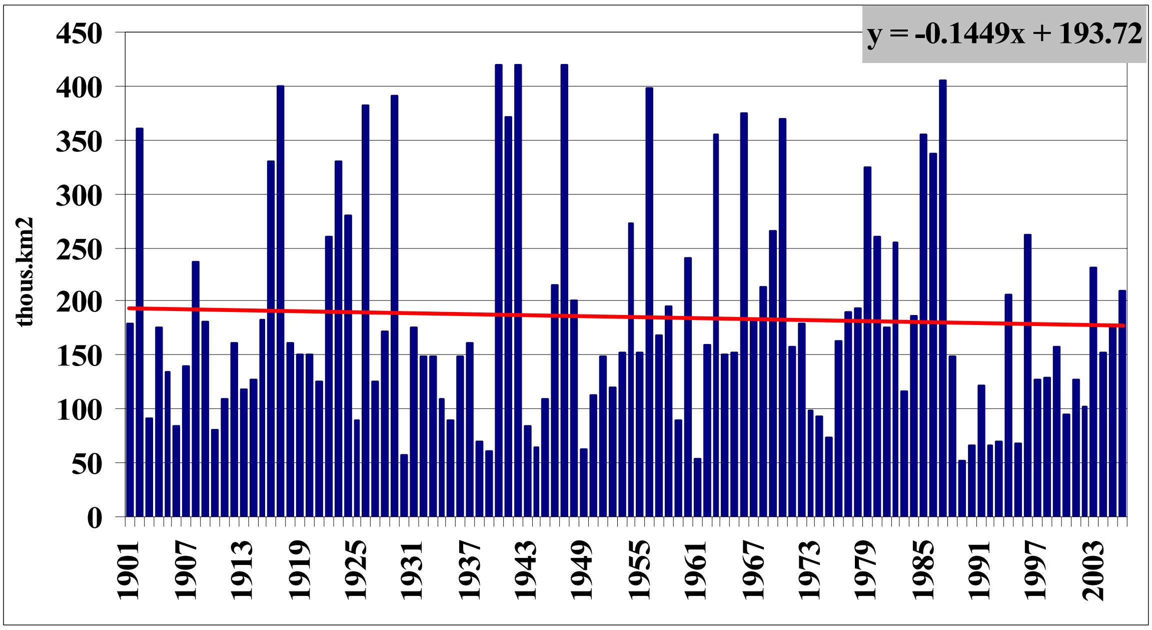

Figure 6. Maximum ice extent in the Baltic Sea and tendencies, 1901-2006

CARGO TURNOVER IN PORTS OF THE GULF OF RIGA

Riga harbor and six small ports (Skulte, Mersrags, Salacgriva, Roja, Lielupe and Engure) are located in the Gulf of Riga. Their main aspects are:

- The Baltic Sea transport;
- Fishing ships basing place;
- Yacht tourism.

Analyzing dynamic of cargo turnover (1996-2006) in Riga harbor and in small ports (data were obtained from Central Statistical Bureau), trend shows strong increasing tendency (Fig.7). Export dominates in the cargo turnover structure.

If the Gulf of Riga in winter period has difficult ice conditions large numbers of ships were turned to the other Baltic Sea ports where shipping conditions were better and it has some impact on Riga's harbor cargo turnover in winter period. Especially negative impacts from difficult ice conditions in the Gulf of Riga have small ports. They can be closed because the small ports fairways are too shallow for icebreaker "Varma".

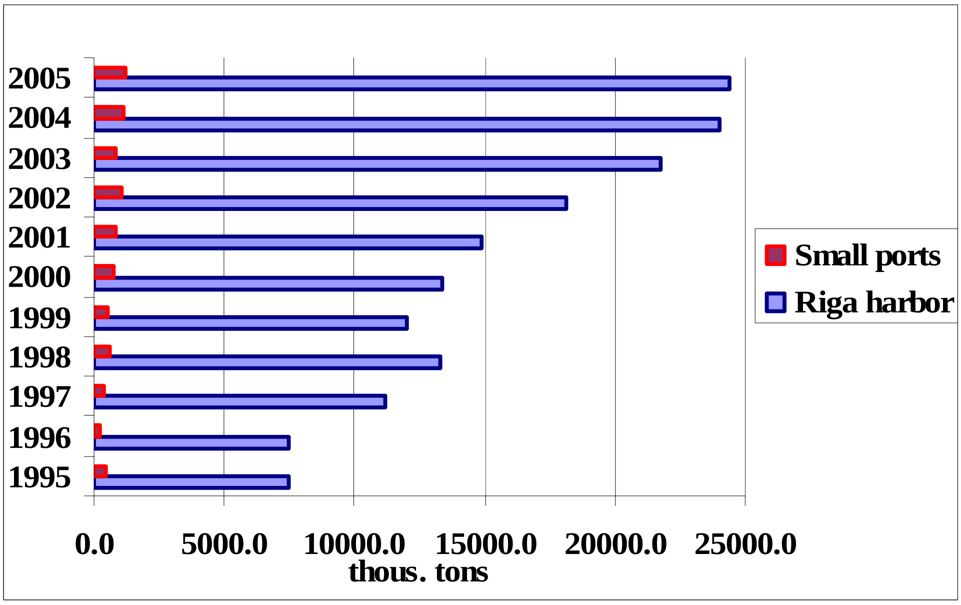

Figure 7. Cargo turnover in Riga harbor and small ports, 1995-2006.

NUMBER OF INCOMING SHIPS TO RIGA'S HARBOR

Navigation to Riga's harbor and small ports can be difficult in the Gulf of Riga even in average winters ice conditions. Navigation to small ports can stop for several days – it depends on severity of winter and availability of icebreaker.

To analyze ice and meteorological parameters impact on winter's navigation 1996-2006, large number of incoming ships to Riga's harbor (Fig.8) was used (data were obtained from Maritime Administration of Latvia).

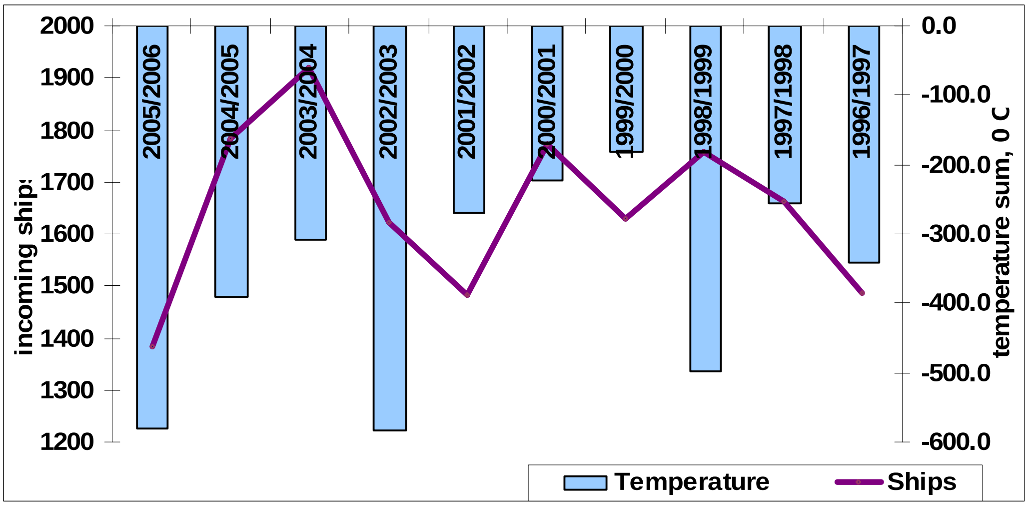

Figure 8. Number of incoming ships to Riga harbor and sum of negative temperatures in Riga in winters 1996-2006.

SOME METEOROLOGICAL PARAMETERS IMPACT ON WINTER NAVIGATION

Wind regime has some impacts on ice conditions especially in coastal areas. Different wind directions in the gulf can make feasible navigation or some of them can cause ice ridging and compacting. Wind directions for different parts of gulf can definite as “good”- ice drift direction from coast, “not good” – ice direction to coast and “neutral” – ice drift along coast (Tab.3).

Table 3. Wind probability (%). Good (+), not good (-), neutral (0) directions (Pastors et al, 1996).

Station	NOV-DEC			JAN-FEB			MAR-APR		
	+	-	0	+	-	0	+	-	0
Rīga	66	15	19	58	23	19	51	35	14
Ainaži	49	29	22	44	27	29	25	51	24
Kolka	20	55	25	26	49	25	24	53	23

Data from this table can be used in planning winter navigation for choosing possible fairway in ice.

Snow in beginning of ice formatting process haste water cooling and ice formatting. After starting ice process formation snow cover not only decelerates ice thickness increasing but also makes addition resistance for navigation. Snow depth maximum on ice usually observe in March – about 15 cm (Pastors et all, 1996).

CONCLUSION

All collected and estimated data are used to understand ice and weather conditions impact on very important sector of economy and give a small look on ice conditions tendency in short time scale.

Ice and weather conditions have significant impact on winter navigation in the Gulf of Riga especially in average and severe winters when icebreaker “Varma” can not provide effective assistance in the gulf’s fairways.

In winters 1996-2006 close connection are shown between maximum of ice extent in the Baltic Sea and sum of negative temperatures in the Gulf of Riga.

In the Gulf of Riga winter severity can be different for its’ regions: milder in southern and western parts and severe in eastern part. This regional ice and weather condition differences also has a big impact on ports’ works.

Analyzing different meteorological parameters for period 1996-2006 was estimated, that in the gulf much more days with ice were observed in March but ice processes started later.

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TOWARDS INTEGRATED SYSTEMS FOR ARCTIC HYDROLOGY – APPLICATIONS AT THE FINNISH HYDROLOGICAL SERVICE

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ABSTRACT

The Arctic-HYDRA programme is a large umbrella for hydrological research in the Arctic. As a multidisciplinary effort, it links hydrology with many other fields of science in geophysical, environmental and social sectors. The spatial focus is the Arctic Ocean drainage basin, but climate and land-atmosphere processes connect Arctic-HYDRA with global studies. From the temporal point of view, the first phases of the programme are forming an IPY (International Polar Year) component between 2007 and 2009. However, the goals of Arctic-HYDRA also aim at long term activities beyond IPY.

This multi dimensional activity calls for integration. There are numerous possibilities to gain synergy: between operational services, between disciplines, between regions, between data controlling systems – and between any combinations of these four components. This paper will focus on the integrated hydrological system concept, where observations, process studies, models, and data systems are effectively linked to produce synergy benefits; this approach is also an important strategic goal of the Arctic-HYDRA. Some developments of the Finnish Hydrological Service have turned out to be successful steps towards integration, and they can offer perspectives to large scale and even Pan-Arctic applications. The Finnish Hydrological Service is located at the Finnish Environment Institute (SYKE), which is both service and research oriented and strongly seeks for integrated solutions.

As a National Hydrological Service, SYKE is operating systems for monitoring, modelling, data processing, and water related geo-information. All of these systems have national coverage and they are mutually integrated at several levels. The first three components – monitoring, modelling and data processing – are also integrated at daily operational level.

The Finnish national hydrological monitoring programme comprises some 1,200 observation sites, which are financed and maintained either by the hydrological service, other research community, or water industry. About 60% of the stations transfer almost real-time data, which means that these data are collected at a maximum frequency of one day.

Hydrological modelling is represented by the Watershed Simulation and Forecasting System (WSFS), which is highly operational and run daily or more frequently. The hydrological model is based on the HBV concept, and it has been further developed at SYKE since 1980s. WSFS makes surface or ground water forecasts for some 1,000 sites, and simulates the basic water balance components at some 6,000 sub catchments using daily time step. The system also generates hydrological maps, automatic flood and snow load warnings and many other products for water resources management, other expert services, and the public. The forecasts and products of WSFS are available on the Internet.

The processing of monitoring data is carried out by a set of data systems. It is composed of a data collection and quality control software, three databases, and a software package for calculation and analyses. The software for data collection and control, and one database (data delivered by the Meteorological Institute) are used by the operator only. One database (for momentary values or separate measurements) and the software package for calculation and analyses are currently for internal use (within the whole Environmental Administration), and the main database is web based and will be released into the Internet use during 2008.

Geo-information databases are currently being developed into more dynamic data systems. The three components are for lakes (56,000 objects), rivers (25,000) and river basins (6,000).

The above basic blocks of the integrated hydrological system interact at several level. The monitoring and data systems continuously transfer almost real-time data into hydrological models, and if models diagnose some data as inconsistent, this feedback is communicated into the data and monitoring systems. Most ground observations are point measurements, while models produce areal values e.g. for precipitation, snow, evaporation and so on. As wide scale models also simulate run-off for high number of sub basins, in-situ measurements are not always needed. On the other hand, real-time observations at important sites keep the status of models correct. Satellite images on snow cover are used in the snow calculations of hydrological models. Good quality geo-information on water resources is essential basic information for hydrological models, and map interfaces have proved to be applicable, informative and user friendly. The list of examples could be continued.

In addition to the operational components (observation, modelling and data systems), the integrated concept includes process studies. Process studies are made for various purposes, but two main objectives and categories can be marked out. On one hand, large scale process studies aim at better modelling at a river basin scale. This work can be considered as model development, and to be successful, it should seek for balance between relevant science aspects and simple, applicable solutions. On the other hand, small scale process studies produce new information on hydrological processes. During the project phase, large and small scale studies are separate processes, but in the long run, they interact. Thus both categories of process studies are linked with operational modelling and furthermore with observation and data systems.

The above discussion means that structures and contents of monitoring programmes should support both modelling and process studies, and vice versa, scientific knowledge should be used to minimise costs of measurement systems. Our experience shows that integration is cost effective – both from the point of view of system operation and scientific relevance.

From the point of view of Arctic-HYDRA, the key question is how to develop national or regional systems into Pan-Arctic. To what extent can local systems be clustered – e.g. by means of internet and user interface? What activities require completely uniform solutions? Probably monitoring systems and process studies are the most independent components, which should be based on common objectives but can be operated on national or corresponding bases. As data systems and related services produce information out of data, they have an "integration" role and they should have an umbrella, a user interface to provide consistent services and products. Modelling will probably be composed of river basin, regional and global/Pan-Arctic models. The current models should probably be mapped and evaluated, and the scientific community should set targets fior Pan-Arctic hydrological models and their integration with atmospheric and Arctic Ocean models. From the hydrological point of view, one of the greatest challenges is to include adequate ground and river system description into large scale models.

PROCESSES AND MORPHOLOGIES OF ICELANDIC GULLIES AND IMPLICATIONS FOR MARS

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ABSTRACT

Iceland provides an excellent natural laboratory for studies of debris flows and other steep slope water-related transport processes. These phenomena are important for the understanding of terrestrial landscape evolution. Basic morphological similarities between Icelandic gullies and controversial, potentially water-related gullies on Mars suggest that Icelandic gullies may also offer insight into the conditions and mechanisms of Martian gully formation. An understanding of how different processes lead to different morphologies in Iceland could identify diagnostic features to help fingerprint the range of processes operating on Mars. Aerial photographs of study sites in Iceland, along with on-site temperature sensors, provide information about the evolution of gully morphology and timescales of gully activity. We use our preliminary observations to investigate the roles of snowmelt, rainfall, and topography in shaping Icelandic steep-slope features.

INTRODUCTION

Debris flows are water-mobilized gravity flows that carry a poorly mixed slurry of rock and sediment downslope. They are one of the major landforms that influence the shape of high-latitude slopes (Åkerman, 1978; Rapp, 1986), and they are ubiquitous in Iceland. Their distinctive signature morphology consists of a chute-like head alcove, a channel (often raised with levees), and a conical debris apron; taken together, these constitute a ‘gully’. Other steep-slope processes active in Iceland include snow avalanches, nivation, gelifluction, rockfall, and fluvial erosion. Many of these processes overlap spatially with debris flows, competing to shape the morphology of gullies.

Previous studies of debris flows in Iceland (e.g. Decaulne and Sæmundsson, 2007) have focused on their natural hazard relevance. We intend to explore the geomorphologic roles of steep-slope processes in Iceland, and we will investigate the potential relevance of these terrestrial processes to gullies and gully-like forms on Mars.

In 2000, Malin and Edgett published images of alcoves, sinuous channels, and debris aprons photographed by the Mars Orbiter Camera (MOC). They championed these gullies as evidence of recent liquid water flowing across the surface of Mars. Substantial controversy ensued, as the planetary science community argued over how the gullies had formed and whether liquid water was involved. Three main end-member hypotheses emerged: liquid water from a subsurface aquifer (Malin and Edgett, 2000; Heldmann, 2007), runoff from melt of snow deposited during a high-obliquity period (Christensen, 2003; Dickson et al., 2007), and dry granular flow (Treiman, 2003). The pitch of the discussion was heightened with Malin et al.'s (2006) announcement of new, bright deposits in two gullies that had appeared since the last set of photographs was taken in 1999.

In the absence of primary data on Martian gullies beyond orbital imagery and spectral analysis, terrestrial analogs have proven to be a valuable resource for testing new ideas about Martian gullies. Previous terrestrial analog studies include Costard et al.'s (2002) work on debris flows in Greenland, Hartmann et al.'s (2003) study of gullies in Iceland, and the work of Head et al. (2007) in the Antarctic Dry Valleys. With its easy accessibility, Iceland offers an excellent opportunity to study a broad range of steep-slope features and to test their viability as Martian analogs (Black and Thorsteinsson, 2008).

Simple debris flows are the Icelandic landform most similar in appearance to the classic Martian gully. Debris flows require ample water, which is in limited supply on Mars, although liquid water runoff at the Martian surface is possible with the assistance of a plugged aquifer (Malin and Edgett, 2000; Heldmann et al., 2007) or a brine *(*Marchant and Head, 2007). But we stress the diversity of steep-slope processes in Iceland, many of which require little or no liquid water to activate. As Åkerman (1978) has pointed out, gully activity on Earth may have fluctuated since the end of the last glaciation. The basic gully structures may be several thousand years old (Rapp, 1987). However, as our results will show, many processes continue to substantially modify gully morphology in the present day. It is therefore important to identify and describe the full range of these ongoing erosional, transport, and depositional processes. An understanding of how different processes lead to different morphologies in Iceland could provide diagnostic features to help identify the range of processes operating on Mars.

With these considerations in mind, the ultimate goals of our investigation are as follows:

- To gain an understanding of the mechanisms of gully formation and their relationship to gully appearance. The basic mechanisms driving debris flows in Iceland have been described by Decaulne and Sæmundsson (2007), and they include: rain on snow, snowmelt induced by a rapid temperature increase (>10ºC in 24 hours), and long-lasting and/or intense rainfall. The morphological expression of these various causes, however, requires further study.

- To analyze gully activity over longer timescales.
- To characterize the evolution of gully morphology.

METHODOLOGY

We combine ongoing field measurements and observations with analysis of aerial photographs provided by Landmælingar Íslands (LMÍ).

The field measurements include snow density, snow temperature, gully-bottom temperature, slope, and the basic dimensions of the gully. Gully-bottom temperatures are obtained with two electronic temperature sensors, one on Ármannsfell and one on the eastern part of the Esja massif. These Starmon-type sensors have an accuracy of ±0.05°C and they take a temperature reading every minute. We have installed them in the bottom of gully channels, in the hope that episodes of meltwater flow in the gullies will leave a temperature signature. We expect that any snowmelt events resulting in top-to-bottom flow should appear as periods of constant near freezing temperature. The goal of our field measurement program is to assemble a record of changes in gully activity and morphology over the course of a full year, thereby illuminating any seasonal dependence.

The aerial photographs (e.g. Figure 2) promise additional insights into rates of gully formation, stages of activity and dysfunction, and areal distribution. They range in scale from roughly 1:20,000 to 1:60,000. Complete coverage of Iceland is available at intervals of roughly 10 years, from 1945 onwards.

RESULTS

Field Observations

Substantial volumes of windblown snow accumulate in the alcoves and channels of Icelandic gullies. Excavation of a gully on Mt. Esja showed that snow depths at the top of the channel were greater than 2.5 meters, even when the adjacent slope was bare. Observation of other Icelandic gullies indicates that this degree of concentration is not unusual. Nivation (snow wash) and focused snow avalanches may be important geomorphic factors as a result (Rapp, 1986). Nivation over millennia may be a primary agent in shaping the scalloped alcoves where Icelandic gullies originate. Snow density and snow temperature profiles from gullies in Southwest Iceland show temperate spring snow packs, with snow responding to air temperature in the top few centimeters but stabilizing at or slightly below zero degrees Celsius throughout the remainder of the profile. Densities range from 0.26 kg/m^3 to 0.554 kg/m^3.

Icelandic gullies vary in cross-section from V-shaped to U-shaped channels. Luckman (1977) has observed that repeated, concentrated snow avalanches may produce U-shaped chutes, which cannot be easily explained by running water. Among the V-shaped channels, there is a wide spread in depth/width ratios. This may prove a fruitful area for additional measurement.

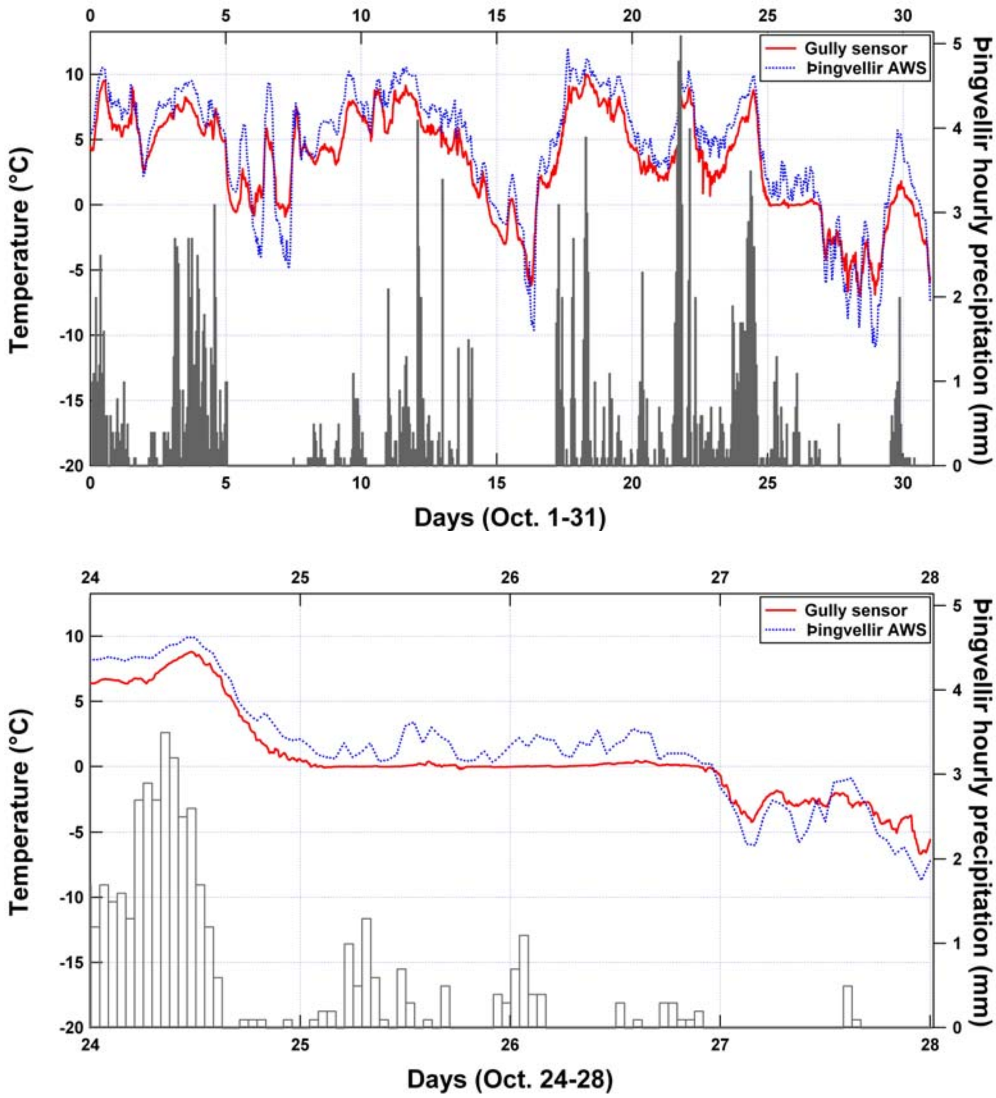

Figure 1. (Top): October temperature measurements from a sensor at Ármannsfell (elev. ~400 m) along with hourly precipitation (gray bars) and temperature at the nearby Þingvellir weather station.

Figure 1. (Bottom): An enlargement of a period of potential gully flow.

Unlike on Mars (Christensen, 2003), older, degraded, and inactive gullies are clearly present in Iceland. Previous authors have used lichen to estimate the elapsed time since the most recent debris flow activity, finding recurrence intervals ranging from 40-500 years in north Sweden (Rapp, 1986).

The temperature sensors reveal one potential episode of flow in the Ármannsfell gully on October 26-27, shown in Figure 1. The three days preceding October 26 were unusually warm, with heavy precipitation (Figure 1). Precipitation on the day preceding the anomaly was especially heavy. We infer that a combination of a warm air mass passing through the area and heavy rain may have triggered snowmelt in the alcove and channel, leading to runoff in the

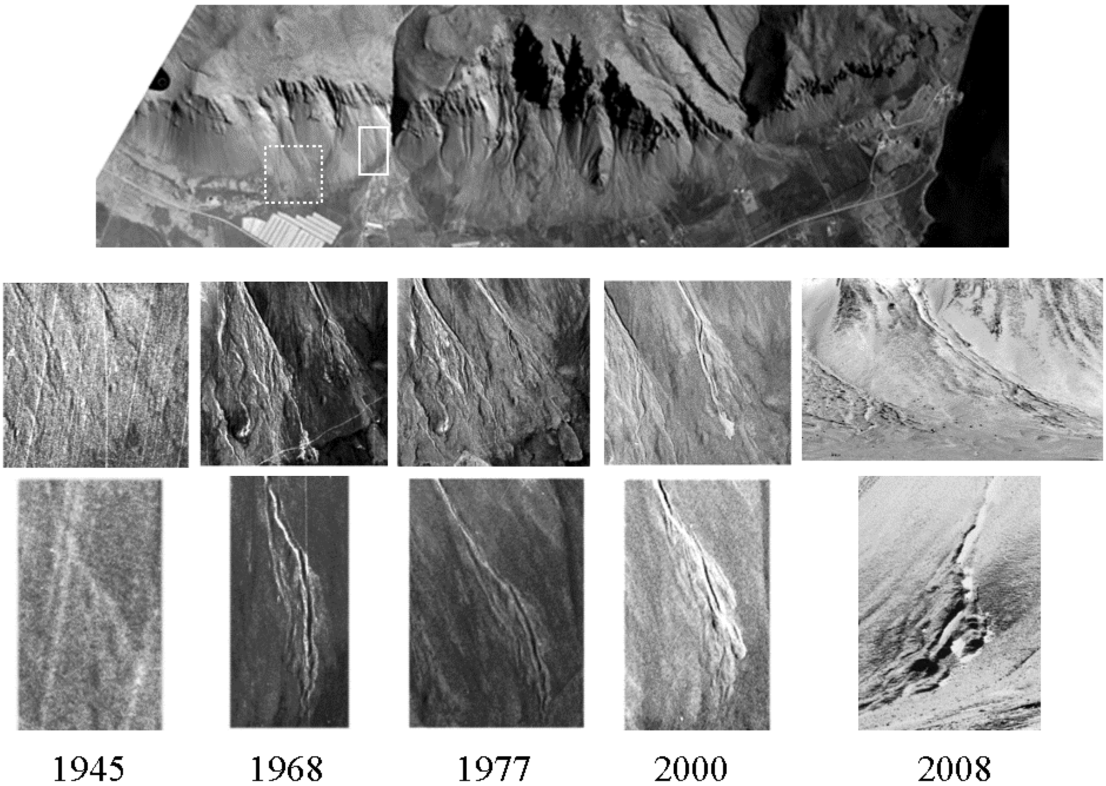

Figure 2. The image at the top provides context of the ~6 km western face of Mt. Esja; the second row and third row present close-ups of fan morphology at two sites in 1945, 1968, 1977, 2000, and 2008. 2008 images are ground-based from January; all other images are aerial, and taken during the summer (courtesy of LMÍ). The gully sites in the second and third rows are indicated by the dashed and solid rectangles, respectively, in the context image. Note the decadal-scale changes in the main channels through the debris aprons, and the snow-filled channels in the 2008 images.

gully. However these results must be viewed with care, as a snow pack around the sensor could also cause zero temperature readings.

The fact that there is rarely only one process at work, even in the same gully, has complicated our efforts to isolate cause and morphological effect.

Orientation does not appear to have a strong effect on large-scale gully distribution in Iceland, although certain faces at specific sites host more gullies than others. Similarly, geology is not a major control either: gullies are found both on hyaloclastic formations and layered basalts. Interestingly, gullies can also occur when there is no upslope drainage network, supporting the role of snow accumulation and melt as a primary driver of gully formation in some cases.

Observations from Aerial Photographs

Analysis of aerial photographs helped us to bracket the timescales of change in Icelandic gullies. On a decadal scale, changes in debris aprons and distal channels were plentiful. Over the course of 50 years, one small new debris flow track appeared on a ~6km face of Mt. Esja (Figure 2, top). We also identified a transient set of intriguing parallel incisions, which appeared and disappeared in tens of years.

Clearly, Icelandic gullies are being actively modified on decadal timescales in the present day. But equally noticeable was the lack of major changes in the landscape of the slope. We did not identify any changes in alcove morphology, and the number of major gullies and channels remained static. We therefore suggest that primary development of gully morphologies occurs on much longer timescales than those for which aerial records are available.

Figure 2 presents close-ups (from 1945, 1968, 1977, 2000, and 2008) of two adjacent sites on Mt. Esja. Changes in the channel paths through the debris aprons, and new debris deposits, are visible at both sites. Whipple and Dunne (1992) argue that debris fan morphology reveals flow rheology: debris-rich flows roughen fan surfaces, and flows with high sediment loads fill in channels, leading to avulsion and the creation of new depositional termini. Time-staggered aerial photographs like those in Figure 2 allow us to analyze the evolution of debris aprons. We observe that individual deposits tend to be rounded and lobate, but as individual deposits accumulate, leading to distal channel avulsion, the overall debris apron may become digitate. This process is illustrated in the bottom sequence of Figure 2.

DISCUSSION

Implications for Iceland

Rapp (1960) and Åkerman (1978) have noted that present rates of talus formation in Spitsbergen seem too low to be consistent with the observed debris fans. They suggest that talus development might be substantially enhanced during periods of increased debris availability. In recent decades glaciers in Iceland have been retreating, possibly due to climatic warming (Sigurðsson et al., 2007). Where valley glaciers retreat to expose slopes, sediments and debris may be released, leading to local increases in the rate of debris apron formation and/or debris flow activity in Iceland. Reported debris flow activity in Iceland has increased in historical time (Decaulne and Sæmundsson, 2007); however it seems likely that this is largely an artefact due to better record-keeping.

Our observations serve to reinforce the diversity of the processes that help to shape Icelandic gullies. Hydrologically, subsurface aquifers do not appear to play a large role in Icelandic gullies. The sources of liquid water are primarily surface runoff and snowmelt.

Figure 3. New bright deposits in a gully in Terra Sirenum on Mars (Malin et al., 2006; HiRISE PSP_004229_1435). Note the digitate terminus.

Implications for Mars

The nature of contemporary gully activity on Mars is one of the major questions in the Mars community today. Terrestrial analogs provide several potential agents as alternatives to liquid water. Snow avalanches, for example, can be facilitated by snow concentration in gullies previously shaped by water.

Pelletier et al. (2008) modelled one of Malin et al.'s (2006) new bright deposits (shown in Figure 3). They found that dry granular flows or very sediment-rich flows produced morphologies more consistent with the observed deposits than the morphologies produced by water-rich flows. Dry granular flows terminated in distributary fingers like those seen in Figure 3, whereas water-rich flows were more strongly controlled by topography and terminated in the same, lowest, location. However, Pelletier et al. (2008) were modelling a single flow event. In Iceland, most gullies appear to host multiple flow episodes. As Figure 2 shows, water-related flows can produce digitate termini; the fingers accrue in separate flows, as deposition blocks old channels and causes avulsion.

Running water, rock glaciers, and gelifluction may all help to erode or remove terrestrial debris aprons. In the absence of any of these processes, debris aprons would likely have been removed during the last glaciation (Rapp, 1986). Thus the maximum age of Icelandic debris aprons may be assumed to be on the order of 10,000 years ago, when the Weischelian glaciation ended. In this context, questions about the age of Martian gullies become vexing. It is difficult to estimate the rate of removal of debris on Mars, especially large debris, but certainly it is much lower than on Earth. Martian gullies do overlay most other young features, including dunes in some cases. However, as our work shows, many other processes can be activated *by the presence of a gully*. The original genesis of Martian gully structures could thus have occurred several tens of millions of years ago, while more recent snow and ice deposition and rockfall may be promoted by the topography of the gully, maintaining a low level of activity.

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SESSION 3: WATER RESOURCES MANAGEMENT UNDER CLIMATE CHANGE

LONG-TERM RECORDS OF PRECIPITATION IN LATVIA

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ABSTRACT

Long-term variability and trends in Latvia's annual, seasonal, monthly and daily precipitation were investigated in this study. Analysis of trends in long-term annual precipitation showed that changes are unevenly distributed over the territory of Latvia. The obtained results indicate that a significant increase in precipitation has occurred in the cold season while the warm period has had a decreasing tendency. The extreme events were studied by the set of climate change indices derived from the longest homogeneous daily precipitation data in Latvia for the meteorological station Riga-University and an increase in heavy precipitation events was found.

INTRODUCTION

Global land mean annual precipitation shows a small, but uncertain, upward trend over the 20th century of approximately 1.1 mm per decade. (Trenberth et al, 2007). However, the records exhibit large inter-decadal variability, and since 1950 global land mean annual precipitation shows an insignificant decrease (Hegerl et al, 2007). Changes in global mean precipitation over land mask large regional variations. There have been marked increase in precipitation in the latter part of the 20th century over northern Europe (Schönwiese and Rapp, 1997) with general decrease southward to the Mediterranean (Piervitali et al, 1998; Romero et al, 1998).

A number of studies have been carried out to investigate whether trends exist in precipitation records (Groisman and Easterling, 1994; Jaagus, 2006). During the last century the amount of precipitation has increased by 12% in the territories that lie between 55^{o}N and 85^{o}N (Folland et al, 2001).

The number of heavy precipitation events over most land regions, even in areas where there has been a decrease in total amount of precipitation, is likely to have increased that is consistent with a warming climate and observed significant increases in atmospheric water vapor. Climatological data show that the most intense precipitation occurs in warm regions (Easterling et al., 2000). Analyses have shown that even without any change in total precipitation, higher temperatures lead to a greater proportion of total precipitation in heavy and very heavy precipitation events (Karl and Trenberth, 2003; Groisman et al., 1999; Katz, 1999).

The aim of the study is to determine long-term variability and trends in the time series of total precipitation, heavy and very heavy precipitation events and to analyze changes in seasonal distribution of precipitation in Latvia.

MATERIALS AND METHODS

The present study is based on annual and mean monthly data series (1925-2006) for 10 meteorological stations obtained from Latvian Environment, Geology and Meteorology Agency. The daily and annual precipitation data of Riga-University station (observations since 1850) was used to investigate the changes in total precipitation over the period of 156 years and to detect extreme precipitation indices (heavy and very heavy precipitation events, highest 1-day precipitation amount).

The seasonal means were calculated as arithmetic means from monthly records. Trends in the precipitation time series were analyzed by using non-parametrical Mann-Kendall test (Libiseller and Grimvall, 2002). The Mann-Kendall test was applied separately to each variable at each site at a significance level of $p \leq 0.01$. The trend was considered as statistically significant if the test statistic was greater than 2 or less than -2. The slope of linear regression was obtained by multiplying the slope with the number of observation years.

RESULTS AND DISCUSSION

Climate in Latvia is affected by its location in the northwest of the Eurasian continent, and by its proximity to the Atlantic Ocean. Because of frequent cyclonic activity over Latvia precipitation occurs about 170-200 days per year. The typical feature of winter weather is the alternation between of cold periods and thaws that last for a few days.

The temporal variability in the trends is seen for the annual precipitation (Fig.1). Overall, the amount of annual precipitation in Latvia has tendency to increase in the remote inland areas.

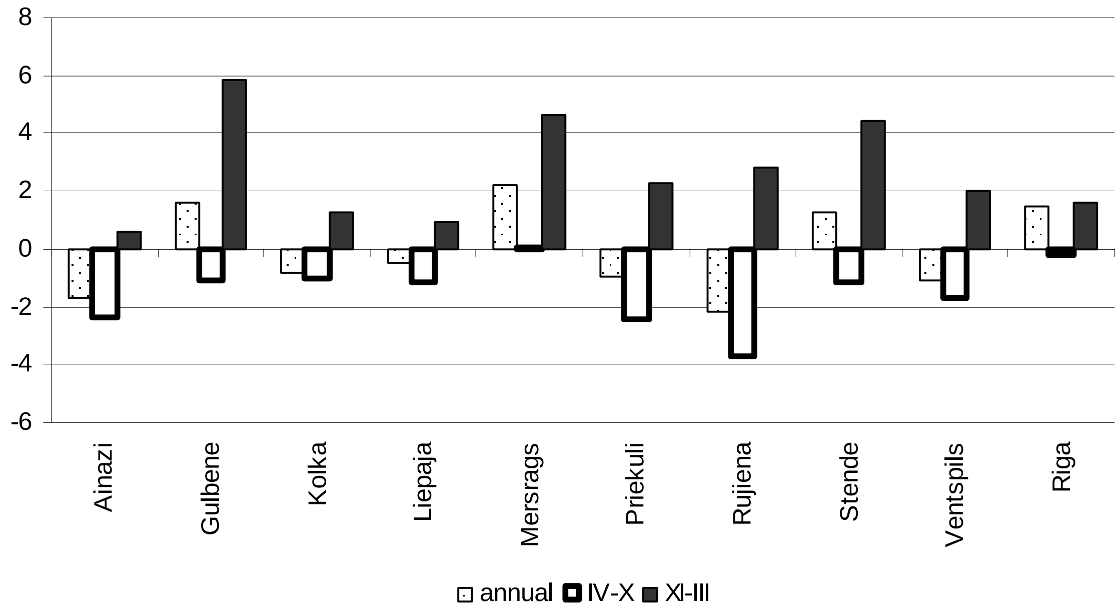

Figure 1. Mann-Kendall test statistics of annual precipitation and precipitation in warm (IV-X) and cold (XI-III) periods (1925-2006).

During the cold half of the year (November - March) statistically significant increase ($p \leq 0.01$ test statistic ≥ 2) in precipitation has been observed at 6 out of 10 stations for the period 1925-2006, but the rest of stations exhibited an increasing tendency. At the same time in the warm period (April - October) precipitation exhibited a decreasing tendency (Fig.1).

Overall, the long-term trend in the annual precipitation in Riga (from 1850) shows a positive sign (Fig. 2). It should be noted that over the entire observation period a large interannual variability and also in some periods interdecadal variability has occurred. The driest period has been between 1858 and 1877 (average annual precipitation 550mm), while the period after 1977 stands out as rather wet (average annual precipitation 700 mm).

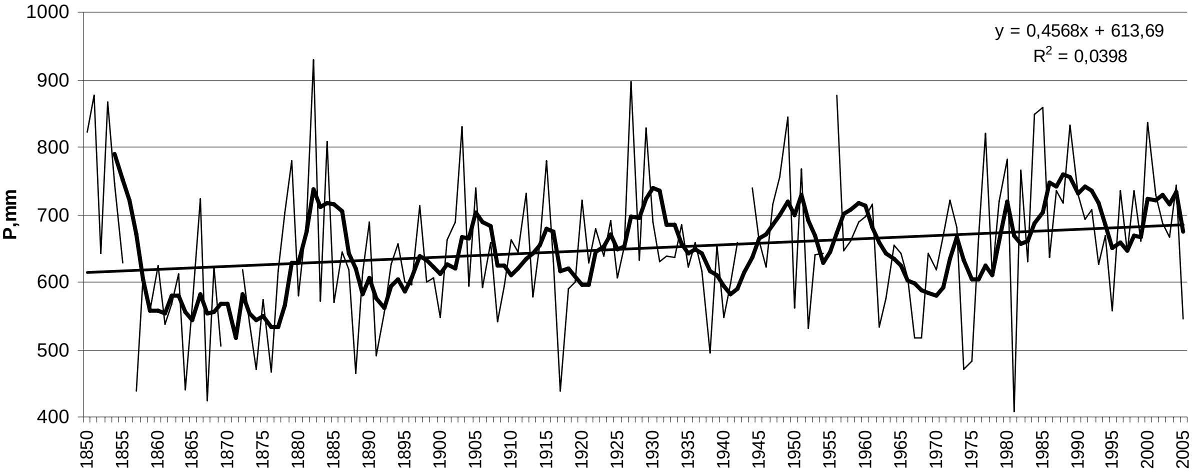

Figure 2. Annual precipitation totals with 5-year smoothed data and trend line for Riga.

Investigation of changes in monthly precipitation series can provide a more detailed overview of the timing of significant changes in annual precipitation. Precipitation has a pronounced tendency to increase in December, January, February, March with a significant decrease in September and July (Table 2).

Table 2. Mann-Kendall test values for the monthly precipitation (1925.-2006.) Statistically significant values (p ≤0.01) are in bold.

Meteorological/ gauge station	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
Ainazi	-0,2	0,2	0,3	-0,6	-1,6	1,1	**-2,1**	-1,7	-1,6	0,0	0,8	1,7
Gulbene	**5,1**	**3,5**	**2,6**	-0,4	0,0	1,4	**-2,6**	-0,9	-1,3	1,4	2,0	**4,6**
Kolka	-0,3	0,3	0,4	-0,3	-1,7	0,9	0,1	-1,2	-1,7	-0,3	0,9	1,2
Liepaja	0,1	0,6	1,8	-1,0	-1,8	1,2	-0,1	-0,1	-1,7	-0,2	0,6	0,5
Mersrags	**3,6**	**3,5**	1,7	0,0	-1,2	0,5	-0,5	-0,2	-1,1	0,3	1,3	**2,9**
Priekuli	**2,2**	1,4	0,8	-0,9	-1,0	-0,9	-1,7	-0,9	-1,6	-0,2	0,4	**2,4**
Rujiena	**2,4**	1,5	0,5	-1,4	-1,7	0,6	**-2,5**	-1,7	**-2,4**	-0,2	0,0	**2,6**
Stende	**2,5**	**2,8**	**2,4**	0,0	-0,8	1,4	-1,9	-1,5	-1,2	0,6	1,0	**3,1**
Ventspils	-0,2	0,2	0,3	-0,6	-1,3	1,5	-0,9	-0,9	**-2,0**	-0,9	0,8	1,7
Riga	0,1	0,6	-0,2	-0,2	-1,0	0,5	-0,1	0,7	-0,4	0,1	0,9	1,9

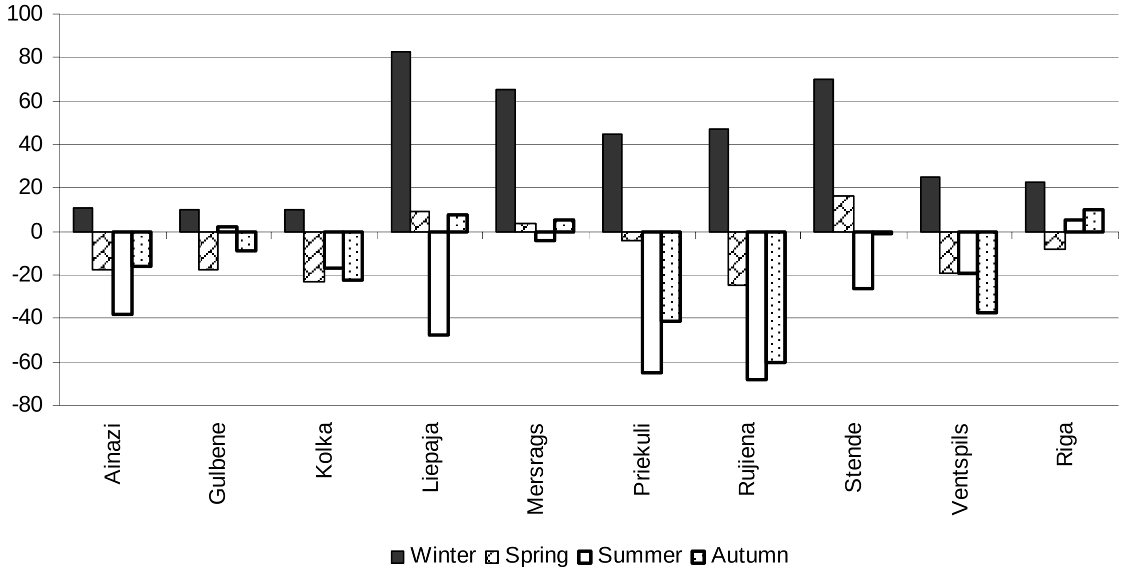

Figure 3. Slope of seasonal precipitation (mm) for period 1925-2006.

The total changes in the amount of precipitation were considered by using slope of linear regression for the 10 stations having 81 years of observation (Fig. 3). It is evident that during winter period the increases in precipitation have occurred, with varying amplitude, in all selected stations. Relatively similar decreasing patterns have been found for summer and autumn seasons, and these are indicative of a decreasing contribution of summer and autumn precipitation to the annual total. The earlier study (Treiliba, 1995) showed that precipitation of cold period in Latvia has become more abundant, and was more evident in those parts, where the prevailing winds and relief fostered ascending of air masses.

Marked precipitation increase was observed in Europe in the first half of the 20 century if compared with period after 1950's. Investigations showed larger increase in winter and autumn seasons and more pronounced in the larger latitudes (Folland et al, 2001). The studies carried out in Estonia detected precipitation increase by 80-180 mm (or 10-25%) in 9 stations during 20 century (Jaagus, 1998; 2006) that is consistent with the tendencies found in Latvia.

The heavy precipitation days (>10 mm) show the positive trend as well as strong interdecadal variability. There have been no clear long-term trends, but a strong interannual, and also interdecadal variability for the very heavy precipitation days (>20 mm) and for highest 1-day precipitation amount (Fig. 4).

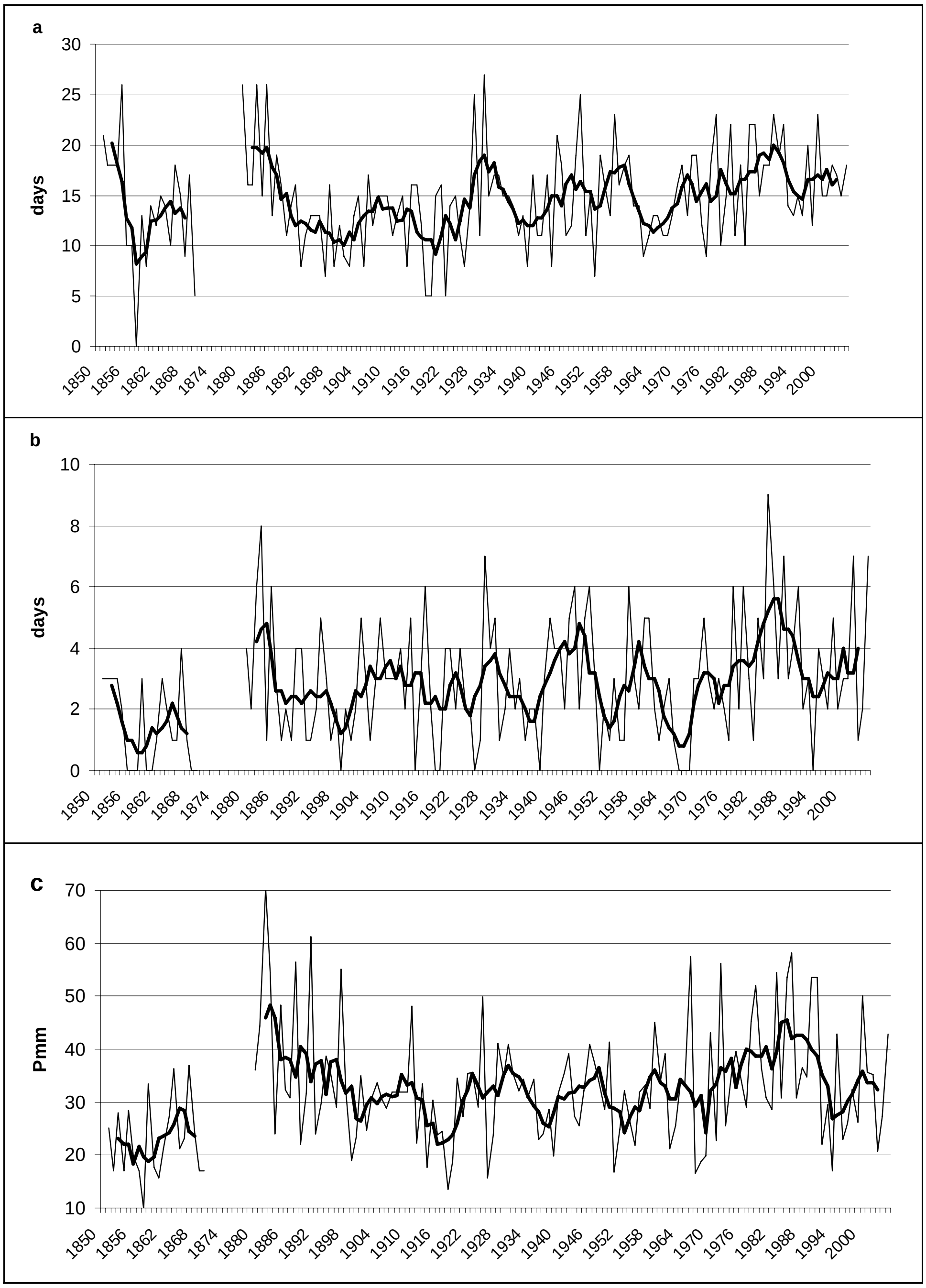

Figure 4. Annual number of heavy (>10mm), very heavy (>20mm) precipitation days and highest 1-day precipitation amount with 5-year smoothed data.

CONCLUSIONS

Analyses of the trends in annual and seasonal precipitation show pronounced intraseasonal and inter-station variations. Overall, increasing trends are evident in precipitation series for the cold period, and monthly precipitation series for most of the stations show increasing trends from December through March. The analysis of the trends in monthly precipitation found a statistically significant decreasing trend only in September and July. Generally, most of meteorological stations exhibited equal tendencies in the given month, but similar physiogeographical regions showed clearly expressed trends neither in annual nor monthly precipitation series and it could be due to fact, that spatial distribution of precipitation is very much a local phenomenon. The time series of precipitation in Riga showed an increase in the number of heavy precipitation events.

ACKNOWLEDGEMENT

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CLIMATE CHANGE AND POLLUTION RISKS RELATED TO INFILTRATION IN PARTIALLY FROZEN SOILS

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ABSTRACT

Large amounts of de-icing chemicals are used in the Northern hemisphere to maintain winter safety on roads and airports every year. At Oslo airport, Gardermoen, Potassium Formate (KF) is used for the de-icing of runways and Propylene Glycol (PG) is the active component of the de-icing chemical used for airplanes. Previous studies have concluded that chemicals can be completely degraded before reaching the ground water levels if there are sufficient retention times. On the motorways in the area, Sodium Chloride (NaCl) is used to keep roads free of ice. Studies of infiltration processes during snowmelt at Gardermoen have revealed a spatio–temporal variable infiltration pattern due to the formation of basal ice on the ground surface. Focused infiltration during snowmelt can reduce the retention time. This has been documented by time-lapse electrical resistivity measurements and modelling. Salt dilution is increased when infiltration takes place uniformly along road-sides. In this paper we suggest a conceptual framework for evaluating risks of focussed releases of contaminants to groundwater or surface waters due to climatic conditions.

INTRODUCTION

Contaminants in melt water from snow cover in the vicinity of airports and roads have received increased attention over the last decade (e.g. Nystén and Suokko, 1998; Øvstedal and Wejden, 2006). The hydrology of partially frozen systems is complicated by the fact that the soil surface may become impermeable due to ground frost and basal ice formation. The unsaturated zone may serve as a filter protecting groundwater from surface pollutants due to bio-geo-chemical processes, hence an evenly distributed infiltration process and long retention times is most ideal. According to

Johnsson and Lundin (1991), Baker and Spaans (1997), Derby and Knighton (1997), French and Van der Zee (1999) and French and Binley (2004), in ltration during snowmelt often occurs as focused recharge in local depressions on the surface because of melt water redistribution on basal ice-cover. This may cause higher velocities in the unsaturated zone than during evenly distributed in ltration on the surface, hence producing less than optimal conditions for degradation. Higher levels of saturation and short retention time in the unsaturated zone will reduce the potential for microbial degradation. Increased redistribution of meltwater can also cause higher concentrations of de-icing chemicals into creeks and constructed infiltration systems along roads.

SITE DESCRIPTION

The studies providing the basis of the conceptual risk map have been conducted on the Gardermoen aquifer, situated 40 km north of Oslo, Norway. It is the largest rain fed unconfined aquifer in Norway. The area is a glacial contact formation with sand and gravels dominating near the ground surface. Hydraulic conductivities (K_s) are in the range 10^{-3} to 10^{-5} m/s. The annual precipitation is about 800 mm/year. In 1996 a new airport access road was opened (Rv35) and the new Oslo airport was opened in 1998, constituting two major de-icing sources in the area. The fate of de-icing chemicals along a road or runway is illustrated in Figure 1.

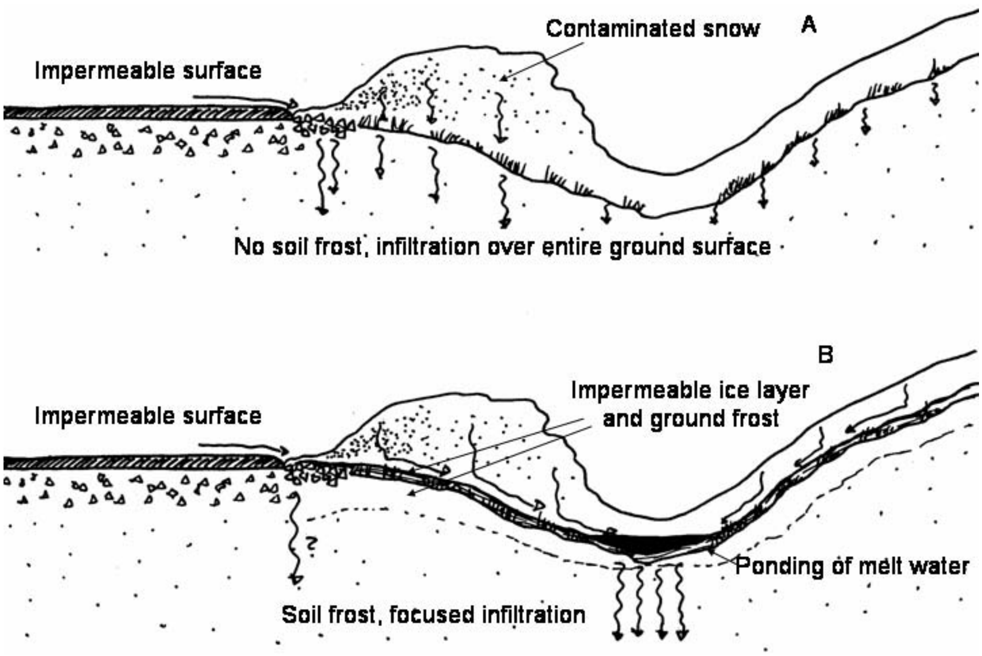

Figure 1. Meltwater flow paths in vicinity of roads or runways A) without soil frost and B) with soil frost.

At the airport Propylene glycol (PG) is used on air planes and Potassium Formate (KF) is used for runways. The de-icing chemicals are easily degradable by bacteria and fungi which are naturally existent in the subsurface. Field experiments and modelling activities documented velocities and degradation rates in the unsaturated zone (French et al., 2001; 2002A; 2002B, Kitterød, 2008). In the case of focussed infiltration chemicals degradation rates may not be sufficient compared to velocities in the unsaturated zone. For optimal degradation conditions it is important that the de-icing chemicals infiltrate evenly over the area. This will ensure sufficient retention time for degradation in the unsaturated zone. An infiltration system for road surface run-off has been constructed along Rv35. A kettle-hole lake, Skånetjern, close to Rv35 have shown a doubling of Chloride concentration since the opening of the road in 1996 (Wike, 2007; Flesjø, 2007).

OBJECTIVE

The issue addressed in this paper concerns how we can relate observed transport mechanisms (including de-icing practices, snow conditions, soil frost and unsaturated flow) of de-icing chemicals from run-ways and roads to surrounding soil and how climatic conditions affect the risk of focussed releases of de-icing chemicals to water sources. This is an important step for evaluating possible consequences of climate change.

FRAMEWORK FOR RISK ESTIMATES

The simple framework for estimating future risks of contamination by de-icing chemicals as a function of climate change consists of three factors or sources of likelihood:

1) Climate conditions:
    - The weather conditions on a specific day, which influence whether de-icing is required or not and
    - weather conditions prior to, or during winter which influence whether the soil becomes frozen or not, whether basal ice forms and at which saturation level the soil freezes at.
2) The need for de-icing, which is strongly linked to the climatic conditions on a specific day, such as snow, rain on frozen roads, represent some of the situations when de-icing may be required.
3) Soil conditions, including soil frost, water contents at freezing, formation of basal ice.

The links between these factors are shown in Figure 2. At this stage we are not estimating actual likelihoods for the different factors but suggesting a framework where such likelihoods can be included for more quantitative estimates of changes in risk due to climate change.

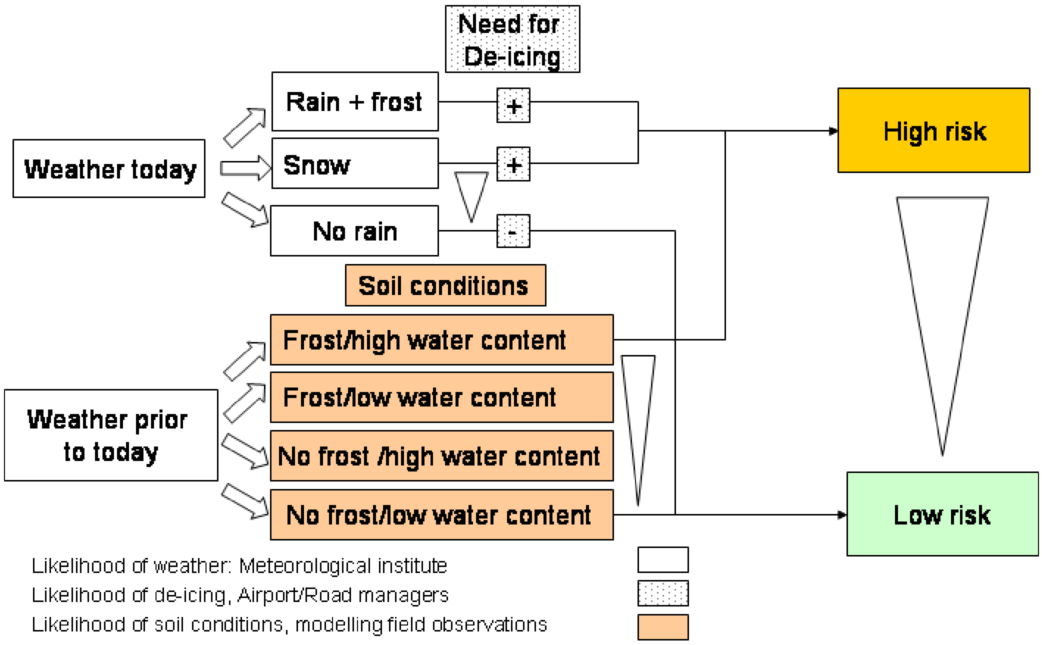

Figure 2. Conceptual framework for risk analysis related to focussed discharge of de-icing chemicals to the groundwater or surface water due to climatic conditions.

In order to evaluate empirically downscaled climate scenarios for the Gardermoen climate station, provided by the Norwegian Meteorological Institute (Engen- Skaugen, 2004), modelled data from the period 1960-90 were compared to observations during the same period. This was done in order to evaluate the uncertainty of estimates of future scenarios for specific locations such as Gardermoen. Only specific winter conditions thought to be important for de-icing practices and soil conditions were considered, such as number of snow days and length of frost periods.

The management practices for de-icing depend on the friction requirement. Since the road (Rv35) and the airport were opened, data on the consumption of de-icing chemicals was registered. In this paper only average values for the winter period, defined here as 01.10 – 30.04 were statistically examined for any correlation with the climatic factors: average air temperature, no. of days with precipitation, accumulated snow fall, number of days with snow and number of days with frost.

Flow and transport in the soil has been monitored by use of field lysimeters (French et al., 2001) and time-lapse electrical resistivity measurements (French et al., 2002; French and Binley, 2004). In the lysimeter experiments de-icing chemicals were added on the surface and water samples at different depths were taken at regular intervals in order to determine velocities and degradation rates. During all experiments climatic conditions, snow accumulation, snowmelt and soil temperature was monitored. Ponding of meltwater in a large depression at the airport was also monitored by manually reading the water level on a measuring stick placed in the relevant area. Within the airport area there is also a regular monitoring programme which includes a large number of boreholes for monitoring groundwater levels and chemical composition of water. In order to estimate risks related to future climate scenarios modelling is necessary. So far flow and degradation of de-icing chemicals was modelled using SUTRA (Voss and Provost, 2003) (French et al., 2001) while TOUGH2 (Preuss, 1991) was used to estimate theoretical breakthrough curves as function of varying boundary conditions (Kitterød, 2008). Flesjø (2007), used SUTRA to model the effect of different infiltration regimes along Rv35; 1) assuming infiltration of salt along all soil surfaces, and 2) assuming impermeable soil along the roads due to basal ice and meltwater redistribution. The unsaturated parts of the models still require development to include the freezing process.

RESULTS AND DISCUSSION

The comparison of observed and modelled downscaled winter conditions for the period 1960-90 show that the combination of temperature and precipitation is rather uncertain for winter conditions. This is not surprising since there is a large sensitivity to whether the temperature is just above or below 0°C. This can be illustrated by the number of days with snow (figure 3) which shows that the modelled data gives a higher value than those observed, while one of the future scenarios looks most similar to the observed values. Hence future scenarios for winter conditions are highly uncertain, and other methods may be required to create scenarios usable as input for modelling studies. The normalised consumption of de-icing chemicals at Oslo airport and along Rv35 during the period 1998-2006 is shown in figure 4.

The correlation between the winter (defined as 01.10-30.04) use of different de-icing chemicals and average air temperature, no. of days with precipitation, accumulated snow fall, number of days with snow and number of days with frost are shown in Figure 5. The figure shows that the strongest correlation between climatic factors and de-icing chemicals exists between Propylene Glycol (PG), which is negatively correlated to average winter air temperature, and positively correlated to the accumulated snow-fall and number of days with frost.

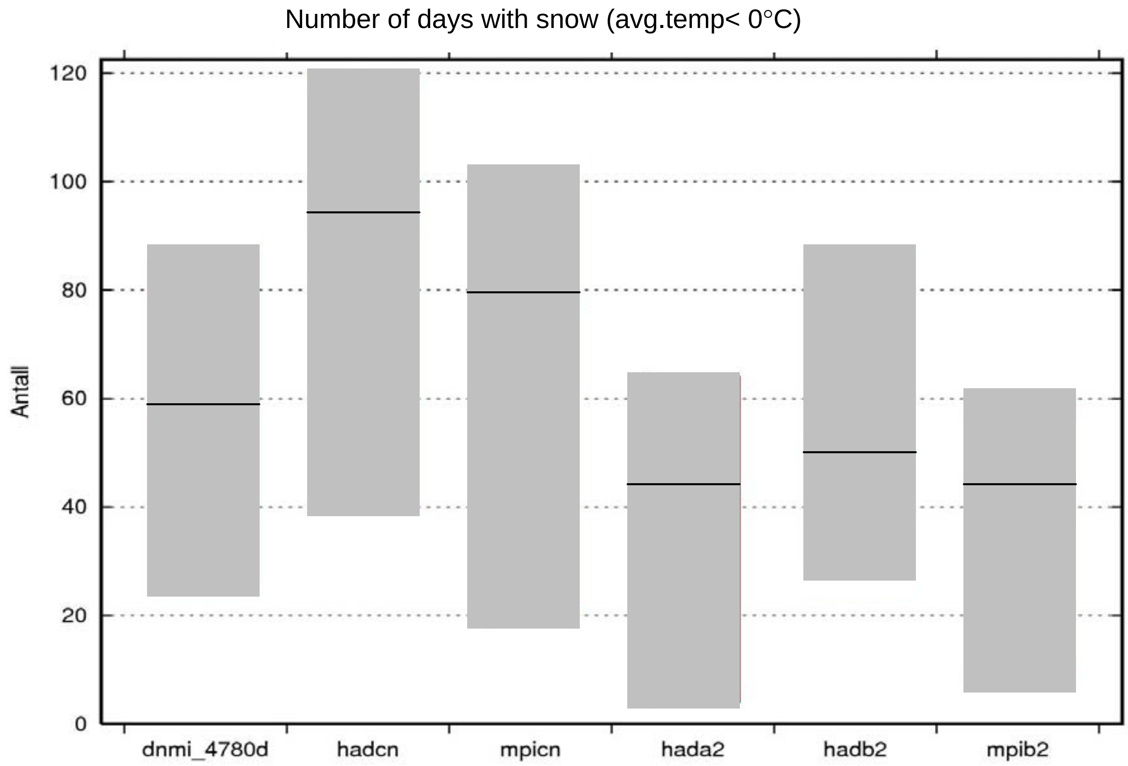

Figure 3. Number of days with snow (Y-axis) and along the X-axis: observed during the period 1960-90 (dnmi_4780d), downscaled scenarios based on the Hadley model (hadcn) and the Max-Planck model (mpicn), and future scenarios for the period 2070-2100 based on the Hadley A2 (hada2), B2 (hadb2) and the Max-Planck B2 (mpib2) global climate models. Centre line indicates mean value, grey box show upper and lower limits.

The de-icing chemicals used for runways and roads are most strongly correlated to each other, of the climatic factors KF shows highest correlation to number of days with precipitation while NaCl is more correlated to accumulated snowfall. Further statistical analysis including more locations and higher time resolution are required for more confidence in these relationships. Based on interviews of salting entrepreneurs, the human factor may also be important (Wike, 2007).

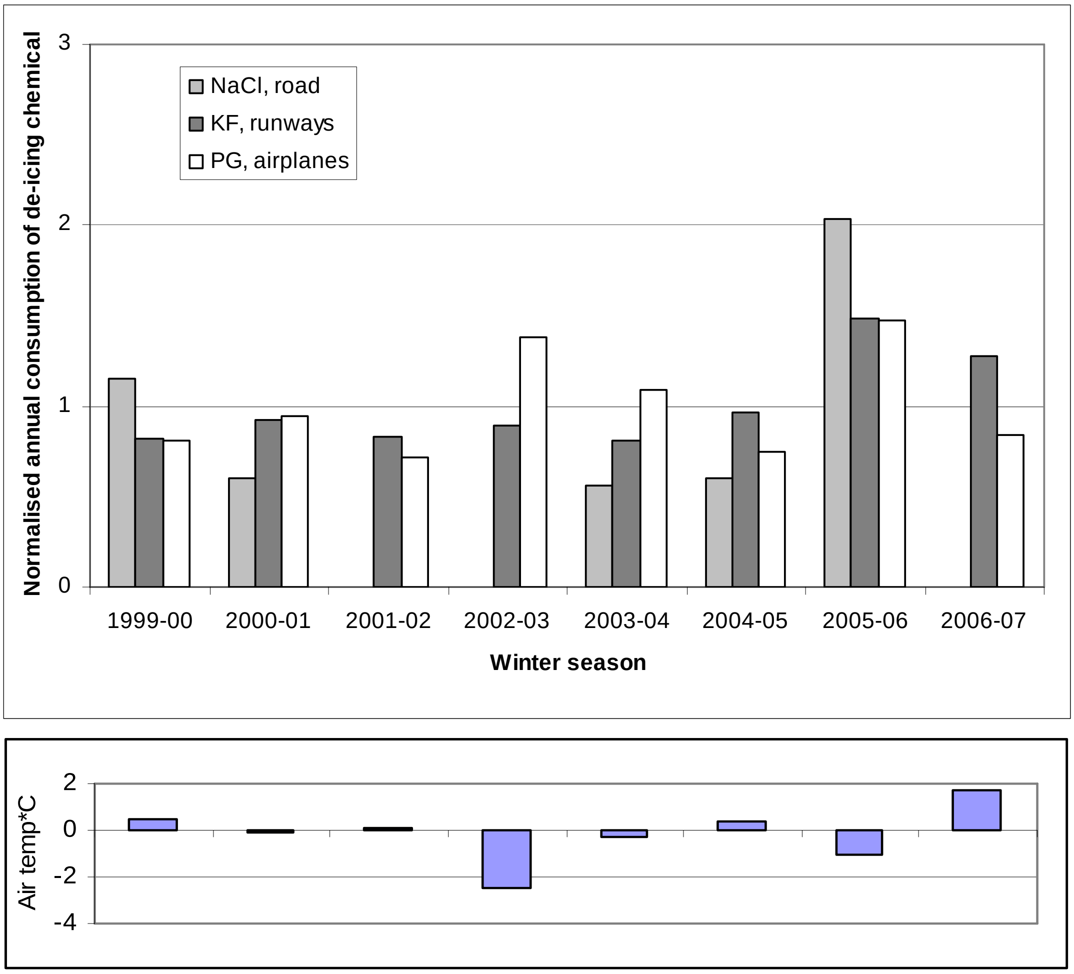

Figure 4. Normalised consumption of de-icing chemicals during the winter period 01.10 – 30.04. on Oslo airport and on Rv35. The average consumption for the different de-icing chemicals are: NaCl:21.6 tonnes/km, KF: 664 m^3 in total, PG:1194 tonnes in total.

The lysimeter experiments conducted during the snowmelts of 1994 and 1995 gave an average vertical displacement of about 7mm per mm of infiltration (French, 1999). Modelling of unsaturated flow using variable boundary conditions indicated retention times in a 4 m thick unsaturated zone between 8 to 12 days for an infiltration rate of 42 mm/day, while an increase to 100 mm/day reduced the retention time to 3 to 5 days (Kitterød, 2008). To illustrate the importance of focussed infiltration in the groundwater zone, figure 5 shows the results of scenario simulations assuming no ice on the surface and one assuming the presence of basal ice and 20% of road runoff is drained through pipes to an infiltration system causing more extreme concentrations than during a more uniform infiltration process.

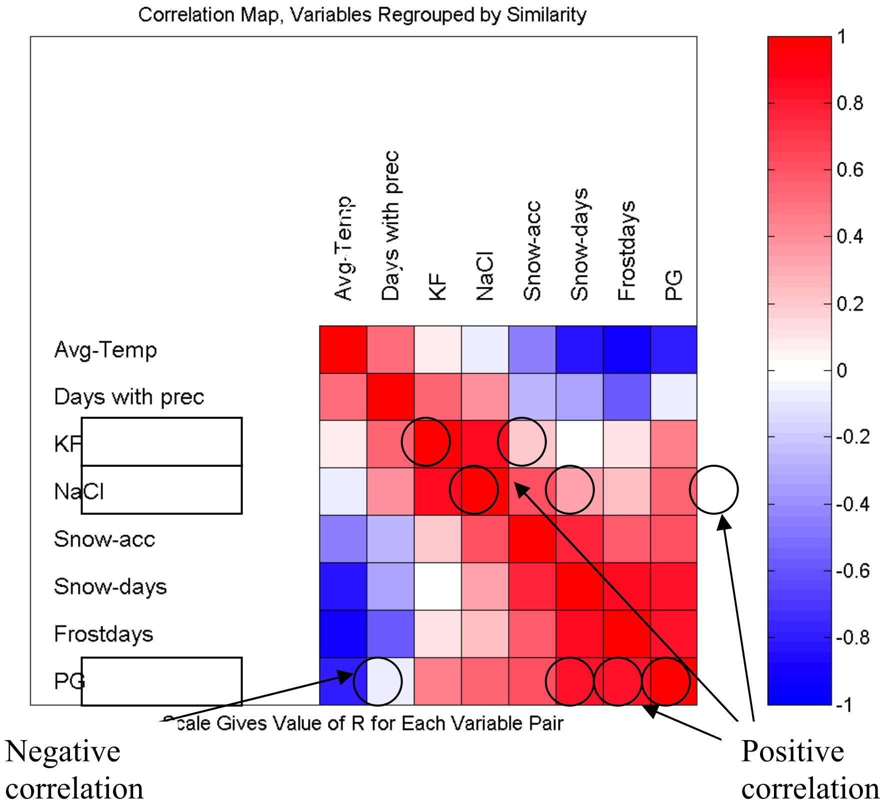

Figure 5. Correlation matrix for average winter climatic factors; average air temperature, no. of days with precipitation, accumulated snow fall, number of days with snow and number of days with frost. Highest correlation for each de-icing chemical indicated by circles.

Based on the present downscaled scenarios for Gardermoen, what could we expect in terms of direction of risk, towards high or low in Figure 2? The average precipitation is expected to increase in all months except June-August in the range of 10-100% (Engen-Skaugen, 2004) this may increase the need for de-icing and the available amount of water for transporting the chemicals. The average temperature in southern Norway during the winter months is expected to increase by about 3°C (Engen-Skaugen, 2004), which most likely will reduce the need for de-icing. With the uncertainty in the downscaled climate scenarios we cannot yet conclude whether the situation goes towards higher or lower risks. The study does however point to which factors are important and which relationships are insufficiently described for a more complete understanding of risk evaluation.

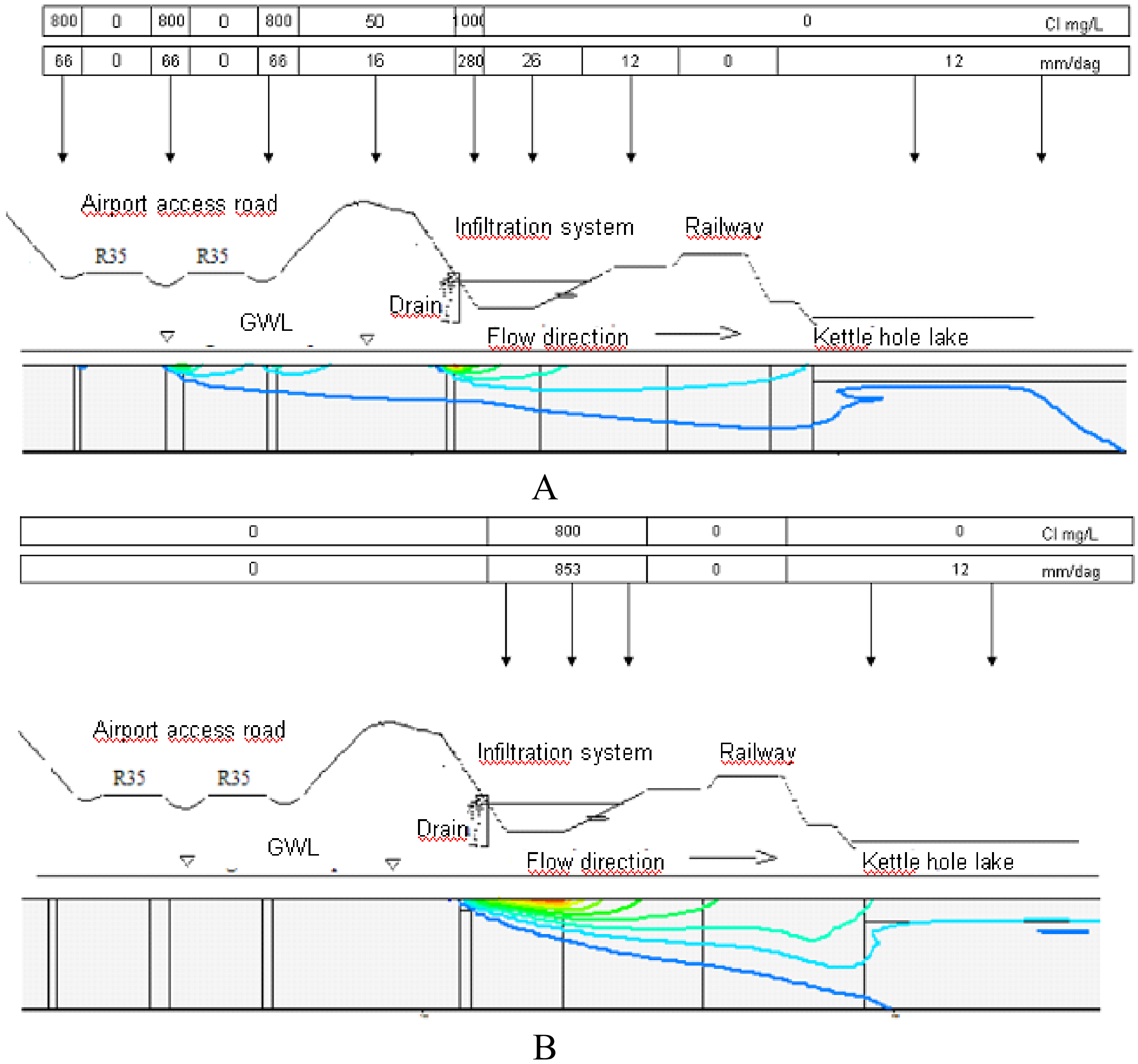

Figure 6. Scenario results of Cl⁻ concentrations in groundwater below a motorway during snowmelt assuming A) no ice on the ground surface B) ice on the ground surface. Boundary conditions of Cl⁻ concentrations and infiltration rates (mm/day) indicated above the vertical sections, based on measured values 2005/2006. Concentrations are relative, with highest concentration near the source area.

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LOW FLOW INDEX MAP FOR NORWAY – INTERACTION USING GIS-SOFTWARE AND ANALYSIS

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ABSTRACT

The aim of the project is to calculate input parameters to regression models to estimate low flow indices for any ungauged catchment in Norway. The first part of the study focused on finding regression equations for homogenous regions in Norway. The regression models require a set of catchment characteristics as input parameters. GIS analyses are defined for all the parameters using national datasets prepared or developed at NVE. The user interacts through a web-application towards a GIS server. This automatically defines the catchment from a point, set by the user in the river network, giving the result from the regression models and the catchment characteristics back to the user.

INTRODUCTION

Estimation of low flow indices at ungauged sites is important for decisions in water resource management. In Norway the amount of planning and construction of small hydro power plants have increased considerably, followed by an extended demand for information about low flow indices in ungauged catchments.

In Norway calculation of common low flow (Engeland et al., 2006), Q_{clf}, is required in the decision-making concerning small hydro power plants. In this project the low flow indices included are Q_{clf} and Q_{95} (the 95 exceedance percentile from the flow duration curve) for the whole year and the summer and winter season separate.

In public administration the need for objective methods are vital and also to use the method which give least uncertain results. The objective of the project is to establish a standard procedure for estimation of low flow indices in ungauged catchments. Regression equations for homogenous regions in Norway are defined (Engeland et. al., 2006). The method is based on several regional regression equations with input parameters calculated using geographic data and GIS analyses tools (ArcGIS with Spatial Analyst extension from ESRI[1]).

The low flow index map in Norway will be a GIS based application from a web user interface. The user indicates, by pointing on an interactive map, where in a river the catchment outlet is located. From this point a catchment is defined using standard ArcGIS tools on a developed flow direction and flow accumulation grid (Fig. 1). GIS analyses are programmed to calculate all catchment and climate characteristics used in the regression equations. The analysis requires national grids and geographic data and includes further development of GIS-routines. The results are estimated low flow indices for the catchment shown on the screen and with export possibilities (Fig. 2).

The presentation will focus on the development of the required geographical datasets and the GIS analyses.

METHOD USED TO GENERATE FLOW DIRECTION AND FLOW ACCUMULATION GRIDS

To be able to generate a catchment from a point, two grids are required, the flow direction and the flow accumulation grid. Both are derived from a digital elevation model. The quality of the results depends on the grid resolution, how the digital elevation model is generated, and which datasets are used as input to the flow accumulation and flow direction grids.

NVE has developed a river network covering Norway and basins with outlet in Norway, at the scale 1:50.000 where all river segments have a direction towards the outlet. From the same scale all lakes with an area above 2500 m^2 are identified with a lake number. For hydrological purposes a new, improved hydrological elevation model is generated for each basin using the following procedure. From the national digital elevation model with resolution 25*25 meters, a point is exported from each cell. These are applied in the topo2raster[2] method using points with elevation, rivers, and lakes with elevation as input parameters. The surfaces are calculated for areas 25*25 km and merged for each basin.

Sinks in this surface raster are filled to remove small imperfections in the data. To make sure that the flow direction follows the rivers, the filled surface is reduced 25 meters where the rivers flow. From this grid, flow direction and flow accumulation grids are derived.

If the river network is correct, this method has proved to give high accuracy catchment boundaries. Trials shows that, for our dataset,

[1] ESRI – Environmental System Research Institute, US.
[2] ESRI ArcGIS Spatial Analyst method. Interpolated a hydrologically correct surface from point, line and polygon data.

catchments derived automatically from a user defined point in the river were very similar to the manually digitized catchments.

CALCULATING CATCHMENT CHARACTERISTICS

The Low flow index map project demands automatically calculated climate and catchment characteristics. This requires availability of some specific national datasets, pre-processing of some datasets and creation of a method for calculation of all the characteristics required in the regression equations. The availability of datasets like a water management system with predefined and identified basins and sub-basins (REGINE), a database with identified lakes, a river network and a runoff map (mean annual runoff) is needed and exists in Norway.

Methods for calculation of land use classes and hypsographic curves are already developed. Calculation of effective lake percentage (the percentage lake in the catchment weighted based on the storage properties of the individual lakes) requires information of the catchment area of each lake in the catchment defined. Combining the lake feature class with the flow accumulation grid, the maximum number of cells draining to each lake was derived. Using the cell size, the catchment area of each lake in the database is calculated and stored in a table with the national lake number as primary key.

Total precipitation and mean temperature grids for each month where made available from the national meteorological institute (met.no). From these, new grids where developed for summer and winter seasons and for the whole year. The average values for the catchment are estimated and stored and used in the regressions equations.

The catchment characteristics describing river length and gradient require the possibility to switch from co-ordinates to Linear references[3] using distance along line features. Defining rivers as Routes[4] allows this possibility.

In the river network all river stretches forming a river or a tributary catchment in REGINE, are identified with the REGINE number code. These rivers are selected and defined as routes in ArcGIS with the source as zero distance. The river length is estimated as the distance from the source of the route to the outlet of the catchment. Together with the maximum elevation retrieved from DTEM25 and the elevation at the outlet, the river gradient is calculated. The 1085gradient, i.e. the river gradient when the 10

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[3] Linear reference - A method for storing geographic data by using a relative position along an already existing line feature; the ability to uniquely identify positions along lines without explicit x,y coordinates. In linear referencing, location is given in terms of a known line feature and a position, or measure, along the feature. Linear referencing is an intuitive way to associate multiple sets of attributes to portions of linear features.

[4] Route - Any line feature, such as a street, highway, river, or pipe, that has a unique identifier.

% lowest part and the 15 % highest part of the river are excluded, is calculated based on using the route distance to find these specific points and their altitude.

Basin length is the distance from the outlet to the most distant point at the catchment boundary. To calculate this automatically the catchment boundaries vertexes are converted to points. Using the ArcGIS function Near[5] the point with the highest value is used as basin length.

CALCULATING LOW FLOW INDICES IN A WEB APPLICATION

The calculation of low flow indices is split into three parts:

1. calculate catchment area – Fig. 1
2. calculate catchment characteristics – Fig. 2
3. calculate low flow indices with catchment characteristics as input.

ArcSDE[6] is used to store both the outlet (point) and calculated catchment. Predefined feature classes for both the point and the catchment are populated with the result of the calculations of catchment characteristics and a universal ID is used as the primary key. Python scripts are developed for calculation of all the parameters. These scripts are sewn together into an ArcGIS model which is published for use on the web with an ArcGIS server. The services enable functionality to be accessible from the web. A web application is then developed to access the services with a map interface used to define a point and showing the resulting catchment boundaries. A tabulated presentation of the generated catchment characteristics is also included in the web application.

The calculations of the low flow indices are performed by feeding the regression models with input variables from the ArcGIS models. A framework to handle the regression coefficients is build on top of an xml structure including the regression equations. This functionality is exposed using standard web services with a SOAP protocol.

[5] Near – Determines the distance from each point in the Input Featire (catchment boundary) to the user point.

[6] Near – Determines the distance from each point in the Input Featire (catchment boundary) to the user point.

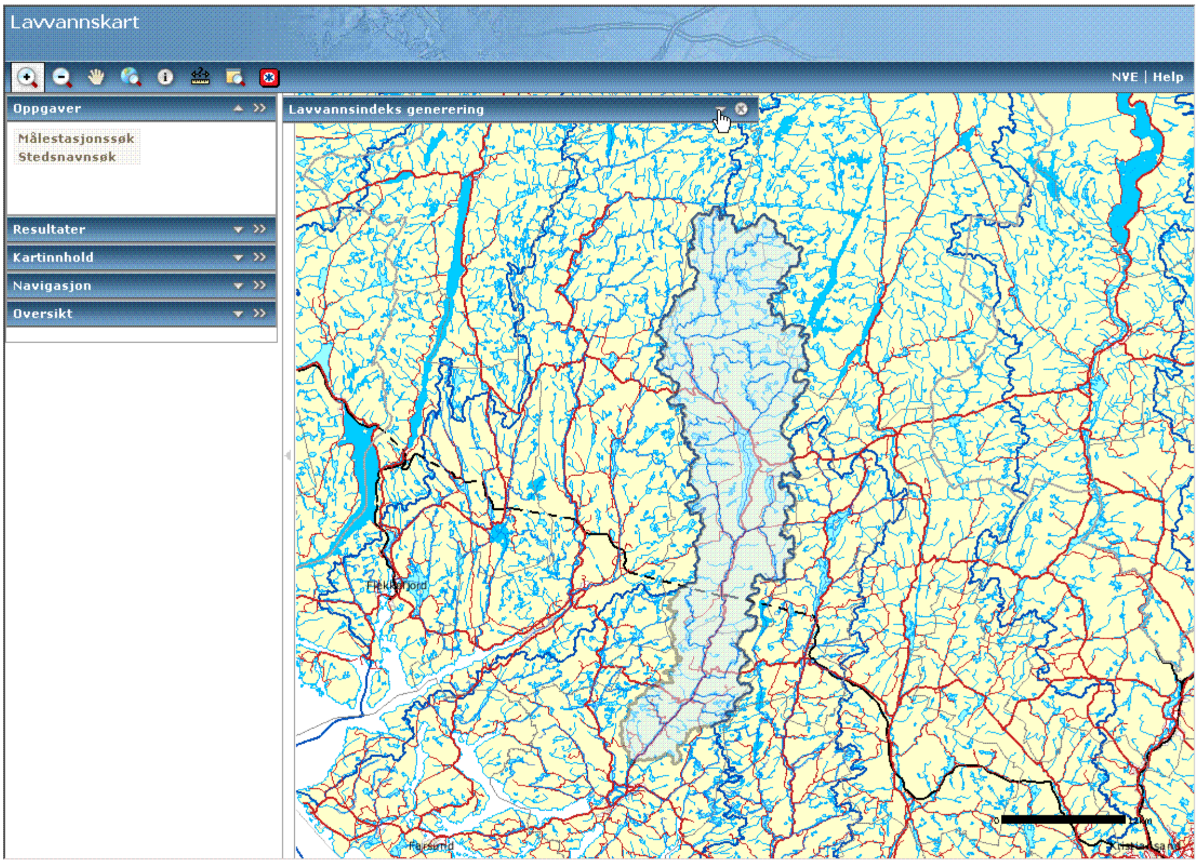

Figure 1. Catchment defined from a user defined point.

CONCLUSION

The results of the GIS analyses are very good. With the presented method to create both the flow direction and the flow accumulation grid, the catchments derived automatically from a user defined point in the river are very similar to the manually digitized catchments. The river network is a main input parameter and the result is dependent on this dataset's correctness. Errors occur and different routines are used to find these by checking the result against manually developed catchments from REGINE.

Our goal is to have 'ready to use' grids and other required datasets covering the whole Norway by the end of 2008. Some of our basins are transnational, having parts of their area in Sweden, Finland or Russia. The finalization will depend of availability of base map data for these areas. The method has not been tested on the large basins and problems in these basins can postpone the completion of the project.

A method to calculate river length and gradient in rivers without a route defined, has not yet been developed.

All the calculations are objective and take less than five minutes. The system increases the efficiency and accuracy in hydrological calculations.

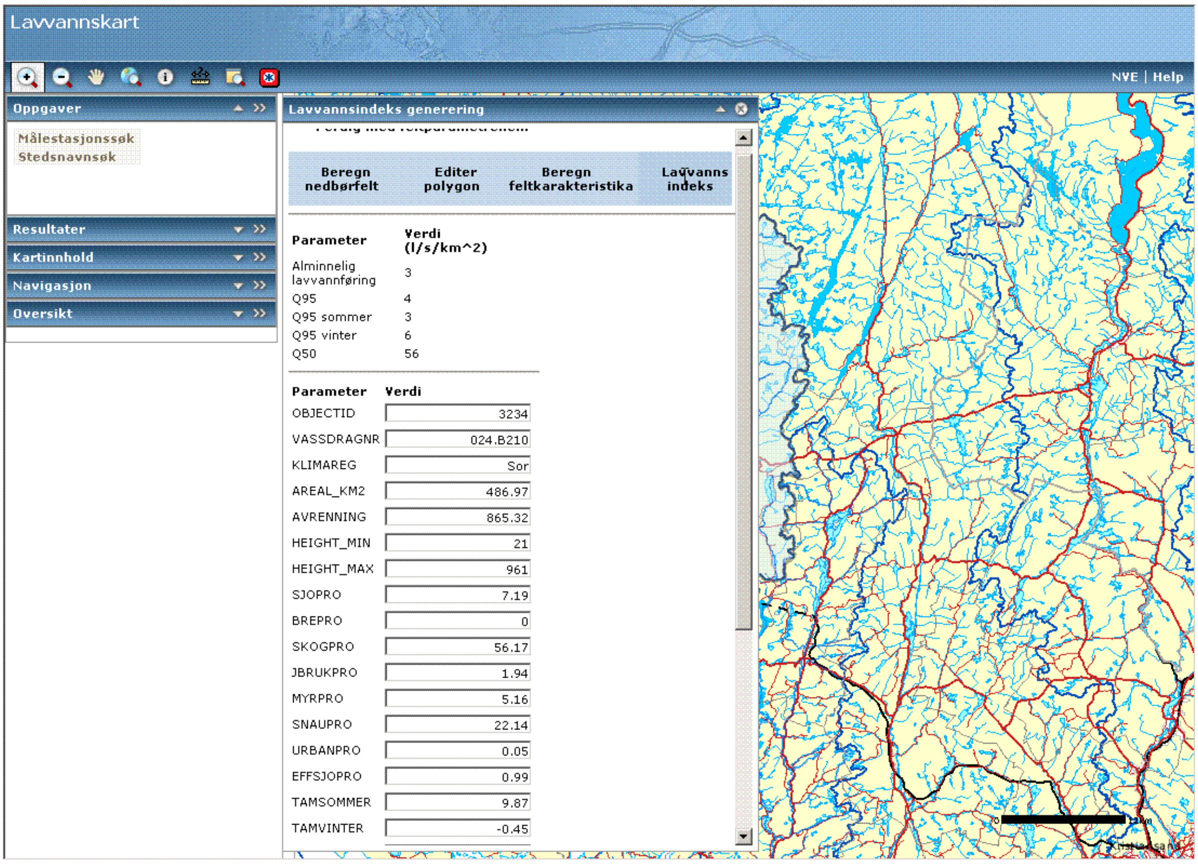

Figure 2. Catchment characteristic and low flow indices calculated for the catchment.

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LONG-TERM CHANGES IN DISCHARGE REGIME IN FINLAND

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ABSTRACT

This paper presents long-term discharge trends and variability for twenty-five time series including both rivers and lake outlets in Finland. The longest discharge time series date back to the mid-1800s. The discharge peak flow usually occurs in the south in April and in the north in June. In northern Finland the maximum flow of the year is always due to snow melt, but in southern Finland summer, autumn and winter high flows are also possible. The Mann-Kendall trend test was applied to study changes in annual, monthly and seasonal mean discharges, maximum and minimum flows and in addition the date of peak flow. The trend analysis showed no changes in mean annual flow in general, but the seasonal distribution of streamflow has been shifted. Winter and spring mean discharges have increased at least in third of the observation sites. The magnitude of increase in winter and spring discharges was 2…10% per decade. Minimum flows have increased at some of unregulated sites and decreased at some of regulated sites. The spring peak has moved earlier at the third of the studied sites with magnitude of 1…3 days per decade.

INTRODUCTION

Water resources are highly dependent of climatic conditions. Run-off regime is affected by both precipitation and temperature changes, as well as changes in radiation balances. Finland belongs to the so called humid zone. A typical feature in Finland is the abundance of water bodies, both lakes and rivers. The water situation may, however, greatly vary from year to year. Within a year, there is a considerable difference between the winter, when the precipitation is stored in the snow cover, and the summer, when a major part of the rainwater evaporates. In the long run, slightly more than half of the precipitation evaporates and little bit less than half flows into the seas from Finland. Discharge gauging is the most precise method of all water balance component measurements. Future climate scenarios predict both droughts and floods to be intensified due to greenhouse effect, therefore it is interesting to examine observed changes in the Finnish streamflow regime hitherto.

There are several earlier studies concerning long-term changes in Finnish discharge regime. Hyvärinen with his colleagues have carried out most of the analysis done in Finland (Hyvärinen and Vehviläinen 1981, Hyvärinen 1988, Hyvärinen and Leppäjärvi 1989, Hyvärinen 1998, 2003). Kuusisto (1992) has also studied long-term runoff from Finland. Finnish streamflow records have also been included in the Nordic runoff studies conducted by Hisdal et al. (1995, 2003, 2004) and Roald (1998). Effects of climate change on water resources in Finland have been presented by e.g. Vehviläinen and Lohvansuu (1991), Hiltunen (1992, 1994), Vehviläinen and Huttunen (1997) just to mention a few.

DATA AND METHODS

The flow regimes in the rivers and lake outlets in Finland were investigated using the records of the Finnish Environment Institute. The typical annual regime, variation and trends of discharge are discussed in this paper. Twenty-five discharge time series, both unregulated and regulated rivers and lake outlets, were examined. Many of the watersheds in Finland are regulated either for water power production or flood mitigation. The longest continuous records date back to the mid-1800s. The lake Saimaa water lever time series and the corresponding discharge time series for the river Vuoksi are available since 1847 and the discharge time series at Muroleenkoski since 1863. Most of the observation series examined in this study started in the 1910s–1930s. The study sites and their locations are presented in Table 1 and Figure 1.

Table 1. Discharge observation period, drainage area (F, km^2), and lake percentage of drainage basin (L, %).

Observation site	Period	F (km^2)	L (%)
Unregulated			
1. Ruunaa (lake outlet)	1931–2004	6 259	13.7
2. Pääjärvi (lake outlet)	1911–2004	1 214	7.1
3. Nilakka, Äyskoski (lake outlet)	1896–2004	2 157	17.9
4. Vantaanjoki, Oulunkylä (river)	1937–2004	1 620	2.8
5. Aurajoki, Hypöistenkoski (river)	1948–2004	351	0.0
6. Kitusjärvi (lake outlet)	1911–2004	546	9.6
7. Muroleenkoski (lake outlet)	1863–2004	6 102	12.2
8. Lestijärvi (lake outlet)	1921–2004	363	21.1
9. Lentua (lake outlet)	1911–2004	2 045	12.7
10. Ounasjoki, Marraskoski (river)	1919–2004	12 303	2.6
11. Tornionjoki, Karunki (river)	1911–2004	39 010	4.7
12. Utsjoki, Patoniva (river)	1963–2004	1 520	2.6
13. Juutuanjoki, Saukkoniva (river)	1921–2004	5 160	4.7
Regulated	Period	F (km^2)	L (%)
14. Pielisjoki, Kaltimo (river)	1911–2004	21 081	13.2
15. Kallavesi, Konnus+Karvio (lake outlet)	1931–2004	16 270	15.3
16. Vuoksi, Tainionkoski (lake outlet)	1847–2004	61 061	20.0
17. Päijänne, Kalkkinen (lake outlet)	1910–2004	26 459	18.9
18. Kymijoki, Anjala (river)	1938–2004	36 275	18.7
19. Kokemäenjoki, Harjavalta (river)	1931–2004	26 117	11.3
20. Kyrönjoki, Skatila (river)	1911–2004	4 833	1.3
21. Lapuanjoki, Keppo (river)	1931–2004	3 949	3.0
22. Kalajoki, Niskakoski (river)	1931–2004	3 065	2.0
23. Oulujärvi, luusua (lake outlet)	1896–2004	19 839	12.8
24. Iijoki, Raasakka (river)	1911–2004	14 191	5.7
25. Kemijoki, Isohaara (river)	1911–2004	50 683	4.3

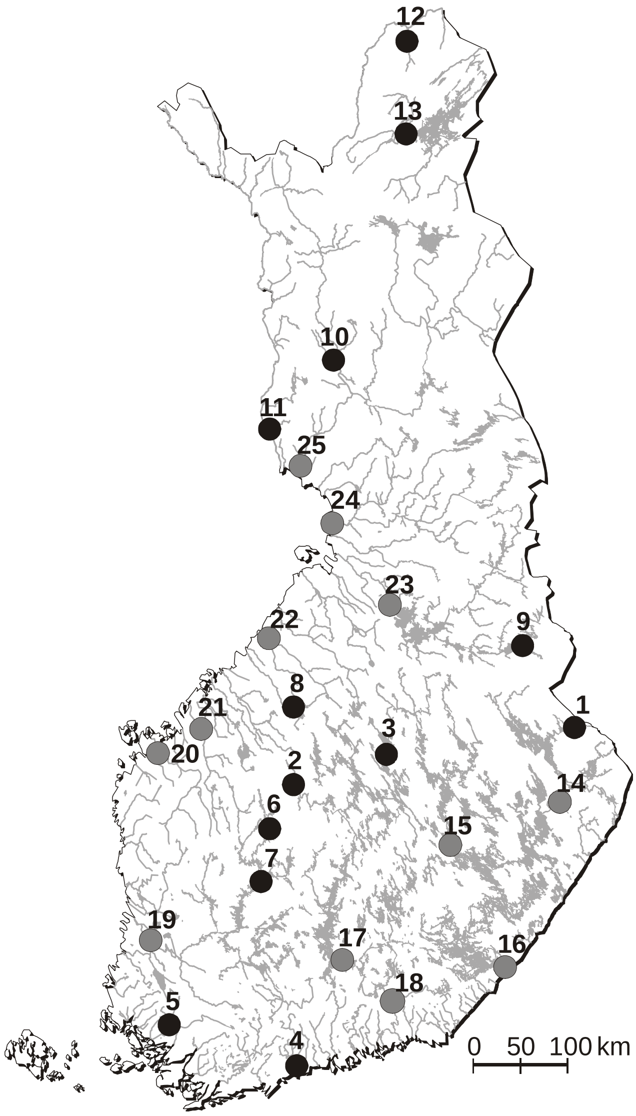

Figure 1. A map of the locations of discharge gauging stations of this study. Black dots present unregulated sites and grey dots regulated sites.

The data were analysed until the year 2004 for the longest available period for all sites. In addition, for unregulated data the period 1961–2004 was studied and for the regulated data both unregulated and regulated time periods. Trend analysis was applied to annual mean discharges (calendar year), monthly mean and seasonal mean discharges (winter: Dec-Jan-Feb, spring: Mar-Apr-May, summer Jun-Jul-Aug, and autumn: Sep-Oct-Nov), annual maximum and minimum flows and date of the peak flow (maximum). Trends were tested statistically with the non-parametric Mann-Kendall trend test. The level of 5% was used for the critical significance. Trend slope of the magnitude was calculated using a non-parametric Sen's slope estimator (Sen 1968). If the data were autocorrelated, the prewhitening procedure presented by Wang and Swail (2001) was applied.

DISCHARGE REGIME IN FINLAND

Precipitation is of course the primary factor that affects the discharge regime. In southern Finland the precipitation is higher than in the north. In the south the growing season is longer, and evaporation is higher. Therefore, the proportion of precipitation that ends up as runoff is higher in the north than in the south. The form of precipitation has a huge effect on the annual discharge regime. In the winter, precipitation is stored as snow. Consequently, water levels and discharges are typically on the lowest level at the end of winter before the snowmelt begins. After that, the highest water levels and discharges are recorded in springtime or in early summer, due to snowmelt. Water levels and discharges usually decrease during summer when the evaporation is normally larger than the precipitation. Sometimes, when the summer is dry and warm, water levels can drop even below the winter minimum. In northern Finland the lowest water levels are normally reached in wintertime, but in the south, especially now that winter discharges have increased, the lowest levels are often recorded in the summer. In autumn, evaporation decreases, and rains raise water levels and discharges. In the small rivers of the southern and western coast, the annual high flow can also occur in autumn, summer, or winter, instead of in spring. In the unregulated rivers in northern Finland the highest flow almost always happens in spring or in summer. When the winter period starts, water levels begin to recede again, because there is no runoff when the soil is frozen and the precipitation falls as snow.

River systems in Finland can be divided into three groups by their discharge regimes. The first group comprises the watersheds of lake regions mostly in southern and central Finland. There are lots of lakes that smooth away annual discharge variations. The river systems of Vuoksi, Kymijoki and Oulujoki, as well as a large part of the Kokemäenjoki river system, are included in this category. Moreover, the waters of the Kuusamo region, which flow into the White Sea, and a few other lake-rich regions, belong to this group. The second group, small and medium-sized river basins with few lakes, is mostly found in the coastal regions along the Gulf of Finland and Gulf of Bothnia. Due to the scarcity of lakes, discharge fluctuations are very rapid. In these rivers, both floods and drought periods are common. The third group includes large rivers in northern Ostrobothnia and Lapland. In these rivers, water flows around the year, even though there are not so many lakes.

The runoff regime of a drainage basin is affected by its area, shape and lake percentage. In large and lake-rich drainage basins, water level and discharge variations are moderate, and small compared to small drainage basins with low lake percentage. In large lakes, water level variations are slower than in small lakes. In large lakes with large drainage areas, the annual maximum water level often occurs as late as in the end of the summer, e.g. in Saimaa, the largest lake in Finland. Water level and discharge fluctuations in rivers with small drainage basins are very rapid compared to those in lakes. The ratio between the mean annual high flow and mean annual low flow can be several hundreds in small rivers, whereas in lake outlets it can be less than ten. The ratio between the mean annual high flow and mean annual flow in outlets of large lakes can be as low as two, while in small river it can be more than fifteen. There are of course variations between the years. The typical variation range of annual mean discharge is 20…40 %. It is largest in small southern rivers and smallest in the north.

TRENDS

At most sites the winter and spring mean discharges have increased at both unregulated and regulated sites (Figures 2 and 3, Tables 2 and 3). In northern Lapland it seems more likely that winter discharges have decreased. The increase of winter discharges focused on late winter and the increase of spring discharges on early spring. Therefore, the rise of winter and spring discharges can be accounted for by the warming in winter and spring and the earlier snowmelt. At some regulated sites the release of water has been increased in winter and in early spring in order to increase the storage capacity for the snow-melt water. This explains the stronger winter and spring discharge trends at some regulated sites. The timing of spring peak flow has moved earlier at about one third of the observation sites. At unregulated sites the change has in most cases been 1…3 days per decade, at regulated sites somewhat more. There is no overall change in the magnitude of spring high flow. At one third of the unregulated sites, summer discharge has increased at least in some months, whereas there has been a decrease in some monthly discharges at slightly less than one half of the regulated sites. The decrease of summer discharges at regulated sites can be at least partly explained by higher water release in winter and spring. At about a half of the unregulated observation sites the low flows have increased, at about a half of the regulated observation sites they have decreased. Increase in the low flow at unregulated sites can be explained by increased discharges in the low flow periods (winter and summer). Decrease in the low flows at regulated sites is explained by zero flows when water gates are shut, but a similar situation is usually not possible in unregulated streams. Annual mean flow and annual high flow did not show statistically significant trends in general, apart from a couple of sites. There were no statistically significant changes in autumn flows in general (Table 2 and 3). Changes in the mean monthly or seasonal discharges were typically some percent of the period mean flow per decade, in most cases not higher than 10%. Trends at the regulated sites were stronger than at the unregulated ones.

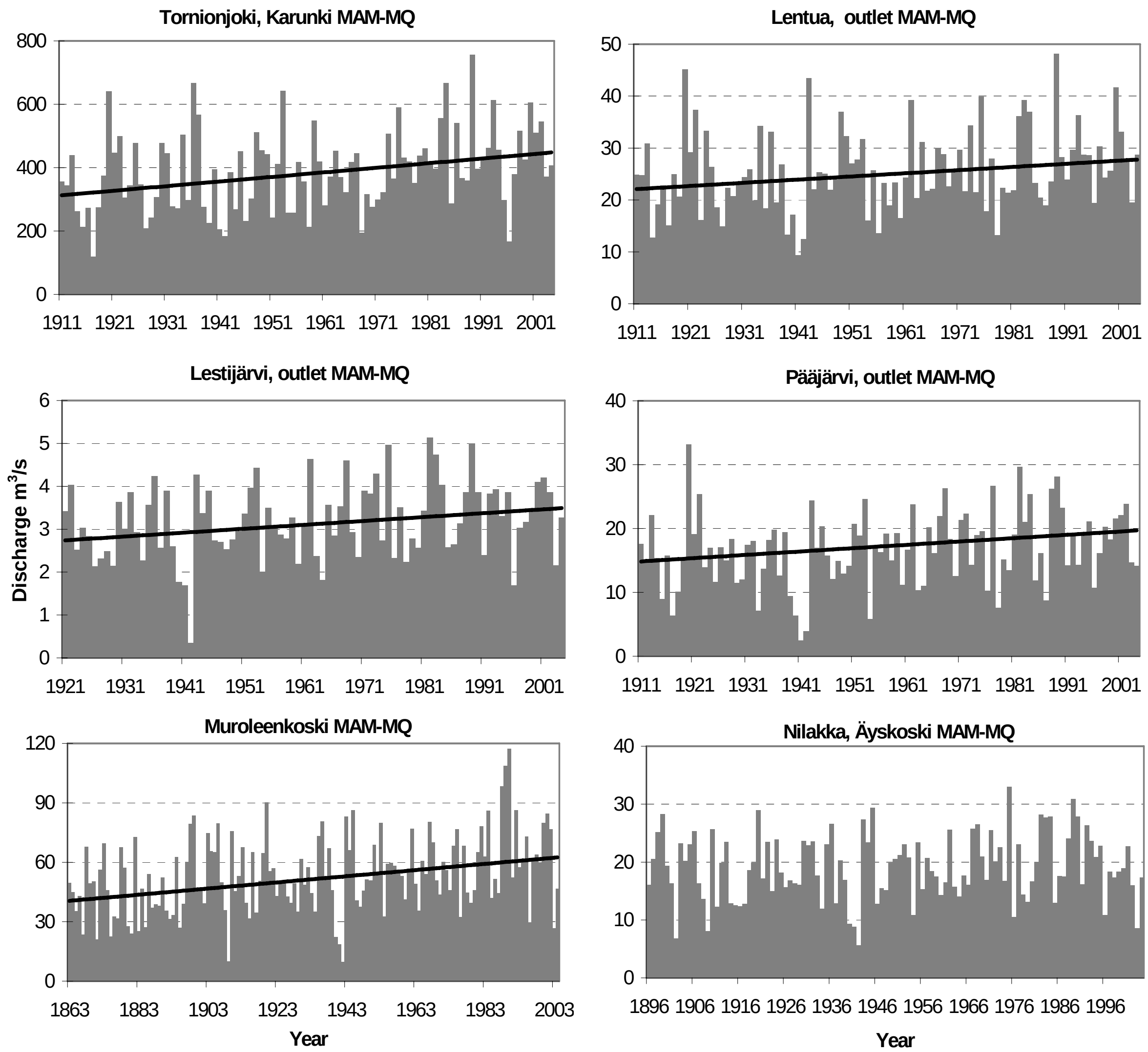

Figure 2. Time series and trends of mean spring (MAM) discharge at some unregulated observation sites. No statistically significant trend for Nilakka.

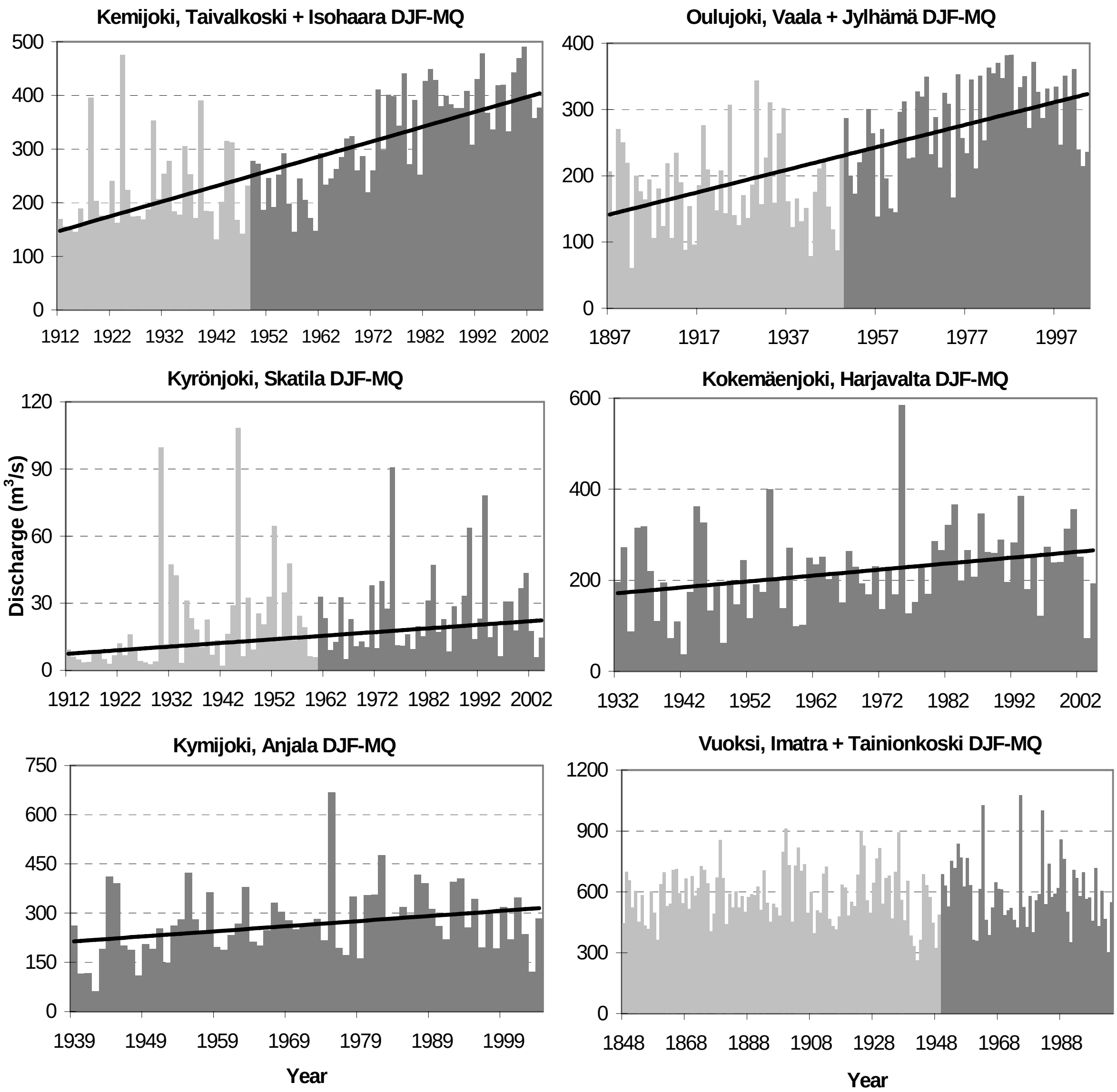

Figure 3. Time series and trends of mean winter (DJF) discharge at some regulated observation sites. Different colours present the unregulated (light grey) and regulated period (dark grey). No statistically significant trend for Vuoksi.

Table 2. Summary of statistically significant seasonal trends, percentage of all unregulated sites (13 sites).

	Winter (DJF)	Spring (MAM)	Summer (JJA)	Autumn (SON)
Positive trend	31%	54%	15%	0%
Negative trend	0%	0%	0%	0%

Table 3. Summary of statistically significant seasonal trends for the whole observation period, percentage of all regulated sites (12 sites).

	Winter (DJF)	Spring (MAM)	Summer (JJA)	Autumn (SON)
Positive trend	42%	50%	0%	0%
Negative trend	0%	0%	17%	8%

COMPARISON WITH OTHER STUDIES

There are a number of earlier studies concerning long-term changes in runoff in Finland and in the Nordic countries. Anterior studies of long-term changes in runoff or discharge regime have shown quite similar patterns as this study. The increase of winter discharges in southern and central Finland was presented first by Hyvärinen and Vehviläinen (1980). Later observations and analyses confirmed these findings (Hyvärinen 1988, Hyvärinen and Leppäjärvi 1989, Hiltunen 1994 and Hyvärinen 1998, 2003). The Nordic studies of trends in runoff regime have revealed considerable differences in different parts of Fennoscandia (Hisdal et al. 1995, 2003, 2004 and Roald 1998). Mean annual discharges have increased especially in some regions in Denmark and Sweden. Positive trends have also been found for Norway and Finland depending on the chosen time period (Hisdal et al. 2004). For the period 1941-2002 statistically significant trends are found for Finland probably because the first year of time period (1941) was the driest ever observed at many places in Finland. In Iceland, annual values of discharge do not show clear trends (Jónsdottir et al. 2005). In Karelia, Northwest Russia, the river runoff has decreased during the 20th century (Filatov 2005).

As mentioned earlier, discharges are naturally highly dependent of precipitation and evaporation. Long-term changes have not been detected for the precipitation time series in Finland (Tuomenvirta 2004), although in the other Nordic countries (Sweden, Norway, Denmark, Iceland) increase has been observed (Hisdal et al. 2003, Jónsdottir et al. 2005). In Karelia, Northwest Russia precipitation has been increased during the 20th century (Filatov et al. 2005). The evaporation time series begin mainly in the late 1950s in Finland, thus such long time series as precipitation and discharge records, are not available for the Class A pan evaporation. However, for the period 1961–1990 no trends were reported by Järvinen and Kuusisto (1995). Neither precipitation nor evaporation show remarkable long-term trends in Finland. Regardless, the changes in the streamflow have been observed in Finland. However, the annual mean flow in unregulated streams has not changed in Finland in general. The main finding is the change in the seasonal distribution of discharge regime.

CONCLUSIONS

Long-term changes in the discharge time series in Finland were investigated in this study. Trend analysis showed that at unregulated sites there has been a statistically significant increase especially in winter and spring mean discharges and minimum discharges at many places. At regulated sites, the discharge regimes have changed quite remarkably: Particularly winter and spring discharges have increased, whereas summer discharges and low flows have decreased at many places. The peak of spring flow has become earlier at one third of all studied observation sites.

There were no overall changes in the annual mean discharges or high flows, neither at unregulated nor regulated sites, apart from a couple of sites. The distinct increase in the winter and spring mean discharges at most sites reflects warming in the winter and spring time as well as earlier springs. These trends are also visible from the climatic time series.

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CLIMATE CHANGE IMPACTS ON THE TOTAL ANNUAL RIVERS' RUNOFF DISTRIBUTION, HIGH AND LOW DISCHARGES IN LATVIA

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ABSTRACT

The paper presents the results of study on the climate change impacts on hydrological regime of nineteen different river basins in Latvia. Analysis of hydrological data series has been done for the periods 1951-2006. Climate change has influenced temporal and spatial distribution of rivers' runoff in Latvia at the turn of century. The present results confirm the basic hypothesis that main tendency in runoff changes are following: decreased in spring flood and as opposite – increased in winter thaw-period.

INTRODUCTION

In Latvia the uneven relief, humid climate, and geological development have developed a comparatively dense network of rivers (mean value is 0.59 km/km^2). Total number of rivers is about 12 500. The mean annual runoff is ~35 km^3, of which only 16 km^3 is formed within the territory of Latvia. The rivers have mixed water feeding: rain, snowmelt water and groundwater. It is typical that the major part of the total annual rivers' runoff has generated in spring season after that followed winter, autumn and summer. Mean annual precipitation varies from 550 to 850 mm. The seasonal distribution of the precipitation favours a wet growing season: about 30% falls in the cold part of the year, and 70% in the warm period (April-October). About 70-75% of precipitation falls in the form of rain, about 15% as now, and the remaining fraction as slush. However, at the turn of century the climate has changed and modified the rivers' hydrological regime as well.

Climate change impacts on rivers' hydrological regime have been extensively studied world-wide in nowadays. Also in Latvia the studies of long-term rivers' runoff and time of ice-break have presented mainly for the large and medium size rivers Kļaviņš et al. (2002), Frisk et al. (2002), Kļaviņš et al. (2004), and rivers' runoff prediction under different climate scenarios Ziverts and Apsite (2005), Rogozova (2006). Theses studies pointed out the river runoff changes in seasonality.

The aim of the study was to analyse the climate change impacts on the total annual rivers' runoff distribution, high and low discharges in Latvia to cover different river basins.

DATA AND METHODS

Data series of nineteen hydrological stations (Fig. 1) have used for the analysis of long-term trends and the distribution of total annual river runoff in percentage by months, seasons and three study periods.

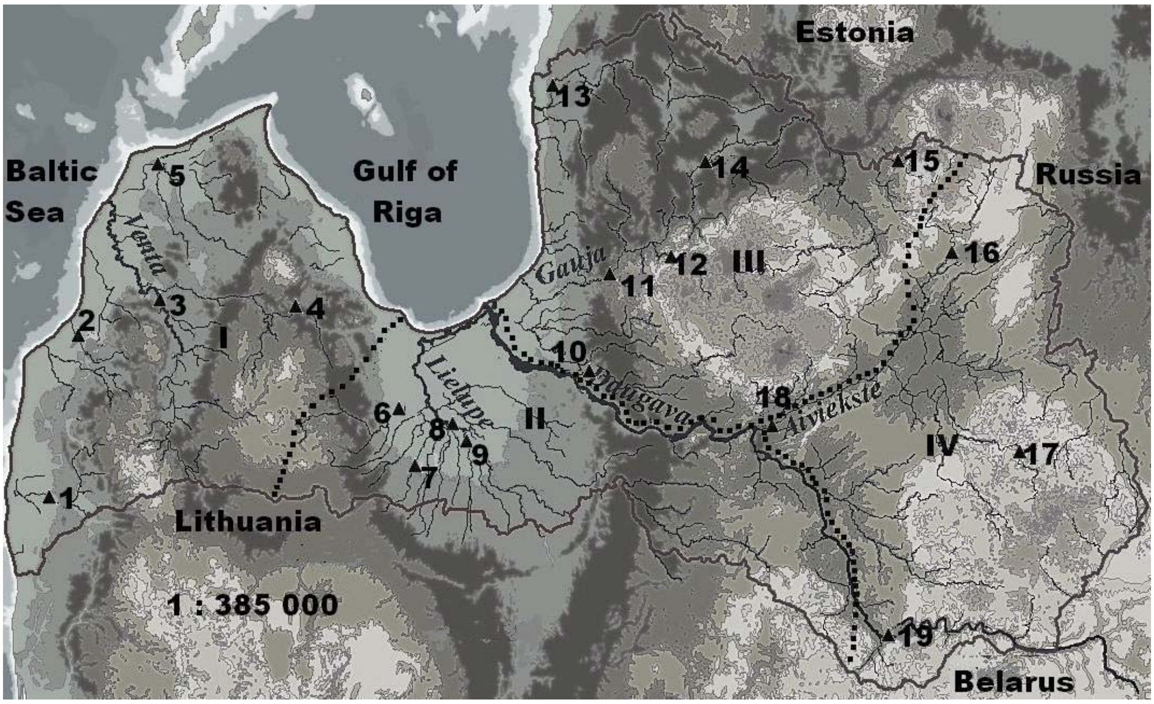

Figure 1. Hydrological districts of Latvia: I – Western, II – Central, III – Northern and IV – Eastern (Glazaceva, 1980). Hydrological stations: 1 – Bārta-Dūkupji, 2 – Rīva-Pievīķi, 3 – Venta-Kuldīga, 4 – Abava-Renda, 5 – Irbe-Vičaki, 6 – Bērze-Baloži, 7 – Svēte-Ūziņi, 8 – Lielupe-Mežotne, 9 – Mūsa-Bauska, 10 – Ogre-Lielpēči, 11 – Gauja-Sigulda, 12 – Amata-Melturi, 13 – Salaca-Lagaste, 14 – Gauja-Valmiera, 15 – Vaidava-Ape, 16 – Pededze-Lietene, 17 – Rēzekne-Griškāni, 18 – Aiviekste-Aiviekste, 19 – Daugava-Daugavpils.

The study periods are following: from 1951 to 1987 (basic – 40-ty years), 1988–2006 (substantial climate change impacts – 19-teen years) and 1951–2006 (all study period). On base of the previous investigation (Kļaviņš et.al., 2007) about the analysis of large-scale atmospheric circulation processes in the Baltic region the study period from 1951 to 2006 was dived into two parts. In this investigation the year 1987 was pointed out as one of the considered climate turning points in a centennial perspective associated with significant changes of climate indicators (winter temperatures, amount of precipitation, etc.).

The low flow periods were defined as a series of the 30-day minimum discharge in could period (Q_{30cold}, November–February) and in warm period (Q_{30warm}, May–October). The high flow period was characterized with maximal discharge of the year (Q_{max}) which mostly was observed in spring flood period (March–April) and sometimes occurred in others seasons of the year. The rations between annual low flow discharge or maximal discharge of the year and annual mean discharge (Q_{30cold}/ $Q_{annual\ mean}$, Q_{30warm}/ $Q_{annual\ mean}$ and $Q_{max}/Q_{annual\ mean}$) were calculated.

Non-parametrical Mann-Kendall test (Lisbiseller and Grimvall, 2002), were used for the data series analysis. The Mann-Kendall test was applied separately to each variable at each site, at a significance level of $p \leq 0.01$. The trend was considered as statistically significant at the 5% level if the test statistic was greater than 2 or less than –2.

The analysis of air temperature and precipitation data series was taken in account for the interpretation of rivers' runoff results. In order to interpret the obtained results, the territory of Latvia was divided into four districts – Western, Central, Northern and Eastern (Glazaceva, 1980). Data source is the Latvian Environment, Geology and Metrology agency.

RESULTS AND DISCUSSIONS

The rivers have typical following hydrograph: two main discharge peaks during the spring snowmelt and in the late autumn during intensive rainfall, and low river discharge in winter and summer. In warm winter the low river discharges can not observed in Western part of Latvia. For this region a comparatively shorter ice cover period can be observed, thus ice break and spring floods begin sooner. There is greater impact on the river discharge regime from meteorological processes occurring over the North Atlantic and Baltic Sea that for others rivers in Latvia, especially to compare with Eastern part.

For this typical rivers' hydrograph corresponds the forty years study period from 1951 to 1987 when climate change impacts were not too proposed on the rivers' runoff. This study period presents the basic distribution of total annual of rivers' runoff: the discharge of rivers during winter reaches 15–30 % of the total annual rivers' runoff, 35–55 % in spring, 2.5–5 % in summer and 5–9 % in autumn. In average the highest percentage in spring runoff has found for the rivers of Central and Easter districts (Table 1). Typically the rivers of Western part preset the highest percentage in winter runoff to compare with others regions. The bulk of total annual runoff came to be in April (16–30 %) and less water discharge – in July and August (2–5 %).

Table 1. Distribution of total annual of rivers' runoff by months and seasons in Latvia (in percentage).

Rydrological district	Month												Season			
	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Winter	Spring	Summer	Autumn
							1951-2006									
Western	11.5	9.8	12.9	16.5	6.0	3.0	2.5	2.8	4.0	6.7	11.6	12.5	33.8	35.5	2.8	7.5
Central	8.8	8.7	16.0	25.3	7.6	3.5	2.9	2.5	3.0	4.8	7.7	9.2	26.7	48.9	3.0	5.2
Northern	7.9	6.9	9.6	23.8	10.7	4.5	3.6	3.9	4.8	6.6	9.0	8.7	23.5	44.1	4.0	6.8
Eastern	6.2	6.1	10.2	25.9	12.9	6.3	4.8	4.1	4.3	5.3	7.0	6.9	19.2	49.0	5.1	5.5
Total in Latvia	6.2	6.1	10.2	25.9	12.9	6.3	4.8	4.1	4.3	5.3	7.0	6.9	19.2	49.0	5.1	5.5
							1951-1987									
Western	9.3	7.5	12.7	18.8	6.3	3.0	2.4	2.9	4.7	7.5	12.0	12.8	29.6	37.8	2.8	8.1
Central	6.2	6.2	16.0	29.4	8.1	3.4	2.6	2.7	3.4	5.5	7.7	8.9	21.4	53.4	2.9	5.5
Northern	6.4	4.8	7.7	26.7	11.7	4.2	3.6	4.0	5.4	7.3	9.3	8.8	20.0	46.1	4.0	7.3
Eastern	5.1	4.3	8.4	29.3	14.1	6.2	4.5	4.2	4.4	5.5	7.0	6.8	16.3	51.9	5.0	5.6
Total in Latvia	5.1	4.3	8.4	29.3	14.1	6.2	4.5	4.2	4.4	5.5	7.0	6.8	16.3	51.9	5.0	5.6
							1988-2006									
Western	14.8	13.5	13.3	12.8	5.6	3.0	2.6	2.6	2.9	5.6	11.1	12.2	40.5	31.7	2.7	6.5
Central	13.5	13.2	15.9	17.8	6.8	3.8	3.4	2.3	2.5	3.5	7.7	9.6	36.3	40.5	3.2	4.6
Northern	8.0	9.0	13.2	20.1	10.7	6.6	5.3	4.1	4.3	4.9	6.8	7.0	24.0	44.1	5.3	5.3
Eastern	10.6	10.7	13.1	18.4	8.9	5.0	3.7	3.6	3.8	5.2	8.5	8.4	29.8	40.4	4.1	5.8
Total in Latvia	9.3	9.9	13.2	19.3	9.8	5.8	4.5	3.9	4.0	5.0	7.7	7.7	26.9	42.3	4.7	5.6
							difference between 1988-2006 and 1951-1987									
Western	5.5	6.0	0.6	-6.0	-0.7	0.0	0.2	-0.4	-1.8	-1.9	-0.8	-0.6	10.9	-6.2	-0.1	-1.5
Central	7.3	6.9	-0.1	-11.5	-1.3	0.4	0.8	-0.3	-0.9	-2.0	0.1	0.6	14.9	-12.9	0.3	-0.9
Northern	2.9	4.7	4.8	-9.2	-3.4	0.4	0.7	-0.1	-0.1	-0.6	-0.2	0.1	7.7	-7.8	0.4	-0.3
Eastern	4.3	5.9	5.4	-8.3	-2.8	0.7	0.1	-0.4	-1.6	-2.1	-0.8	-0.4	9.7	-5.7	0.1	-1.5
Total in Latvia	6.6	7.3	5.4	-12.0	-6.1	-1.6	-0.8	-1.0	-1.0	-0.7	1.5	2.4	16.3	-12.7	-1.1	-0.1

The study period from 1988 to 2006 presents the distribution of total annual rivers' runoff under climate change impacts during last nineteen years. The distribution of total annual of rivers' runoff is following: 19–45 in winter, 28–40 % in spring, 2–5.5 % in summer and 4–7 % in autumn. The results show that rivers' runoff remarkably increase in winter (average 8-15%), decrease in spring (average 6-13%), slate decrease in autumn (~1%) and unsubstantial changes in summer to compare with study period 1951–1987. Theses seasonal changes in total annual river runoff could substantially observed in Central and also in Western parts of Latvia due to occurring warmer winters at the turn of century (Table 1). The analysis of monthly results shows (Fig. 2) that in Latvia river runoff was increased in January and February (average 3-7%) and decreased in April (average 6-12%). Among the regions the most substantial changes in river runoff could found for the Central part - in January and February increase by ~7% and in April decrease by ~12%. In the Eastern part the climate is more continental to compare with others parts of Latvia. Therefore, the changes in river runoff reflected more gradually – increasing in January, February and March and decreasing in April and May.

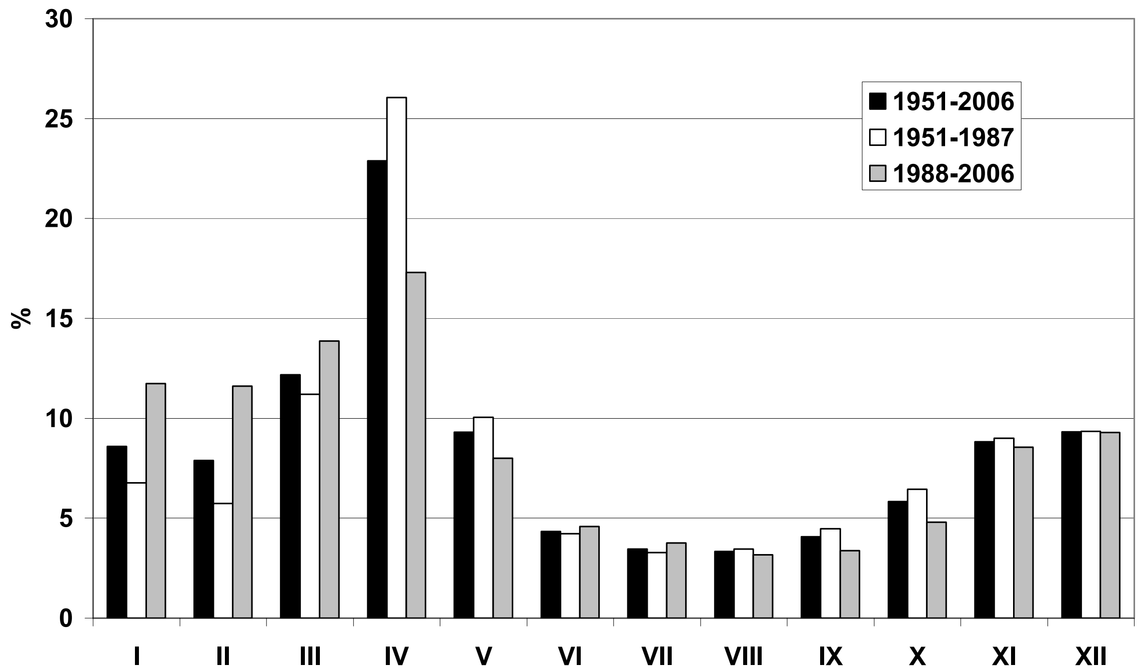

Figure 2. The distribution of total annual of rivers' runoff in average percentage totally in Latvia by studies periods.

The results of monthly discharge for the all study period (1951-2006) with the Mann-Kendall test produce the same conclusions as we discussed above: statistically significant (at significant level of $p \leq 0.01$) upward trend of rivers' runoff in mostly all hydrological stations in January and February and downward trend in 10 of 18 hydrological stations in April (Table 2). In generally statistically insignificant log-term trends came for the summer and autumn months. In Easter and Northern parts proposed increase of river runoff could be see for February and March, and decrease - in April (especially in Northern part). Moreover in Central and Western parts river runoff increased more significant in January and February.

Generally known is the fact that the changes of river hydrological regime are determined with climate changes. In Latvia the results of previous investigations on log-term changes of air temperature and precipitation approved the common tendencies of log-term changes on rivers' runoff. For instance, Lizuma et. al. (2007) has analyzed long-term trend (1950-2003) of seasonal temperature changes. According to the Mann-Kendall test criteria for period demonstrates a statistically significant increase of temperature in all the 22 meteorological stations. The highest increases in average air temperature were recorded in spring (March, April and May) and early winter (November and December). In the later half of winter and in summer the trend was less pronounced. During autumn seasons there was no change in air temperature in Latvia. Also the mean annual maximum temperature was increased more rapidly in April and May while the minimum temperature was increased more rapidly in winter season.

Briede and Lizuma (2007) have concluded that in study period from 1950 to 2003 annual and seasonal precipitation showed significant variations in trends between stations. They have found out that in 13 of 24 metrological stations an overall increasing trend (at significant level of $p \leq 0.01$) was evident in precipitation series for the cold period. Statistically significant upward trends for most of the stations were obtained for January, March and June. A corresponding statistically significant downward monthly trend was obtained only for September. The maximal diurnal sum of the monthly precipitation has also increased in the cold period of the year. They have concluded that North Atlantic Oscillation seems to play an important role in the precipitation regime in Latvia. Particularly, it explains significant proportion of winter precipitation variability in the territory.

In this study log-time changes of the low and high flow periods were analysed from 1951 to 2006. The Mann-Kendall test showed statistical significant upward trend for the 30-day minimum discharge (Q_{30cold}) of could period in 99% of studied hydrological stations and downward trend – for the maximal discharge of the year (Q_{max}) in 12 of 19 hydrological stations (Table 2). Significant downward/upward trends of the 30-day minimum discharge (Q_{30warm}) of warm period were found only in 5 hydrological stations.

Similar results were obtained to analyze ration between annual low flow discharge (Q_{30cold} or Q_{30warm}) and annual mean discharge ($Q_{annual\ mean}$) among tree study periods (Table 3). The ration Q_{30cold} /$Q_{annual\ mean}$ was increased for all studied rivers in the period 1988-2006 to compare with study period 1951-1987.

Table 2. The results of Mann-Kendall test for monthly discharge, annual low, high and mean discharges (1951-2006). Trend is statistically significant at the 5% level (p≤0.01) in bold.

Hydrological station	Month												Q_{30cold}	Q_{30warm}	Q_{max}	$Q_{annual\ mean}$
	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII				
Western district																
Venta-Kuldīga	**2.41**	**2.39**	1.36	**-2.62**	-0.54	**2.51**	**2.93**	0.62	0.05	0.08	0.30	0.42	**2.74**	**2.46**	**-2.40**	0.51
Abava-Renda	**2.79**	**2.67**	0.54	-0.62	0.43	**3.90**	**2.44**	**1.70**	0.81	0.27	0.41	0.54	**2.04**	**2.12**	-1.87	1.13
Irbe-Vičaki	1.40	1.45	1.97	-0.92	-0.99	0.99	**2.49**	0.52	-0.06	0.13	1.25	0.03	**2.42**	1.77	-1.61	1.96
Rīva-Pievīķi	**2.25**	**2.08**	0.76	-0.83	-0.52	0.86	0.58	0.31	-0.55	-0.51	-0.26	-0.02	**2.63**	-0.25	-0.11	1.91
Bšrta-Dūkupji	**1.87**	**2.14**	1.70	**-1.82**	-0.75	1.16	1.74	0.45	-0.92	0.19	0.61	0.50	**2.21**	0.49	-1.47	0.78
Central district																
Lielupe-Mežotne	**3.07**	**3.00**	1.66	**-2.68**	-0.96	**2.83**	**4.21**	**2.34**	1.71	0.64	0.76	1.65	**3.34**	**3.63**	**-2.60**	0.30
Mūsa-Bauska	**2.33**	**2.57**	1.64	**-2.35**	-0.40	1.65	1.69	0.01	-0.11	0.36	0.90	1.39	**2.06**	-1.80	-1.46	0.54
Svēte-Ūziņi	**2.24**	**2.21**	0.98	**-2.71**	-1.02	1.22	1.77	-0.01	-0.23	0.21	1.78	1.46	**2.36**	0.46	**-4.00**	-1.23
Bērze-Baloži	**2.24**	**2.65**	1.34	-1.44	-0.17	1.94	1.25	-0.27	-0.46	-0.61	-0.27	0.34	**3.39**	0.74	**-3.56**	-0.39
Nothern district																
Gauja-Sigulda	**2.86**	**2.91**	**3.13**	**-2.09**	-0.87	1.35	1.15	-0.56	-1.29	-1.71	0.08	0.23	**2.61**	-0.59	-1.53	0.49
Gauja-Valmiera													**3.29**	-0.74	**-2.32**	0.85
Amata-Melturi	1.79	**3.00**	**2.39**	**-2.28**	-0.59	0.38	0.00	-1.61	-2.01	-1.07	0.16	-0.16	**2.57**	**-2.74**	**-3.22**	0.13
Vaidava-Ape	**2.30**	**2.90**	**2.90**	**-2.23**	-1.12	**2.15**	1.42	-0.47	-0.60	-1.00	0.38	0.65	**2.79**	1.12	**-3.75**	1.01
Ogre-Lielpēči	**2.62**	**2.28**	**2.60**	**-2.65**	-0.85	1.58	1.31	-0.98	-1.51	-1.24	0.43	0.31	**2.57**	-0.95	**-3.19**	1.06
Salaca-Lagaste	**2.68**	**2.74**	**2.97**	-0.81	-1.02	1.22	1.48	1.08	-0.41	-0.80	0.14	0.46	**3.38**	1.69	-0.54	1.58
Eastern district																
Daugava-Daugavpils	1.56	**2.57**	**3.02**	-1.80	**-1.69**	-0.19	1.44	1.03	0.81	0.37	0.83	1.11	**2.54**	0.88	**-2.57**	0.98
Aiviekste-Aiviekste	**2.34**	**2.43**	**2.54**	-1.94	**-3.12**	-0.95	1.19	0.71	0.40	0.47	0.66	0.87	**2.61**	1.17	**-2.70**	0.27
Rēzekne-Griškāni	**2.26**	**2.54**	**2.60**	-1.59	1.65	**2.79**	**2.92**	1.44	1.23	0.75	0.85	0.99	**2.45**	**2.58**	**-2.19**	1.93
Pededze-Litene	**2.81**	**2.75**	**2.21**	**-2.03**	-1.46	1.56	0.69	-0.71	-0.63	-0.50	0.13	0.69	1.61	-0.06	**-2.31**	0.26

The higher ration was fond for the Western and Central part (about 0.8) and lover ration – for the Northern and Eastern, respectively 0.73 and 0.7. The ration Q_{30max} /$Q_{annual\ mean}$ was decreased also for all studied rivers in the period 1988-2006. The higher difference of this ration between studied period 1988-2006 (ration 6.96) and 1951-1987 (ration 11.66) was obtained for the rivers of Central part, than following Northern, Eastern and Western. The calculated ration Q_{30warm} /$Q_{annual\ mean}$ wasn't show substantial changes between studied periods 1988-2006 and 1951-1987.

Table 3. The rations between annual low or high flow discharge and annual mean discharge.

Study period	Q_{30cold}	Q_{30warm}	Q_{max}	$Q_{annual\ mean}$
Western district				
1951-2006	0.63	0.21	6.04	1.00
1951-1987	0.53	0.22	6.30	0.98
1988-2006	0.82	0.21	5.54	1.04
Central district				
1951-2006	0.54	0.18	10.07	1.00
1951-1987	0.40	0.18	11.66	1.00
1988-2006	0.79	0.17	6.96	1.00
Nothern district				
1951-2006	0.51	0.25	7.96	1.00
1951-1987	0.40	0.26	8.48	0.98
1988-2006	0.73	0.22	6.95	1.04
Eastern district				
1951-2006	0.50	0.31	6.53	1.00
1951-1987	0.40	0.31	6.98	0.96
1988-2006	0.70	0.33	5.65	1.09
Total in Latvia				
1951-2006	0.55	0.24	7.65	1.00
1951-1987	0.43	0.24	8.36	0.98
1988-2006	0.76	0.23	6.28	1.04

It could be concluded that the rivers' 30-day minimum discharge of could period has typical upward trend and the maximal discharge of the year – typical downward trend in the last twenty years (Fig. 3 and Fig. 4). Also these results have confirmed the previous obtained results on changes of long-term runoff regime of Latvian rivers in time and space.

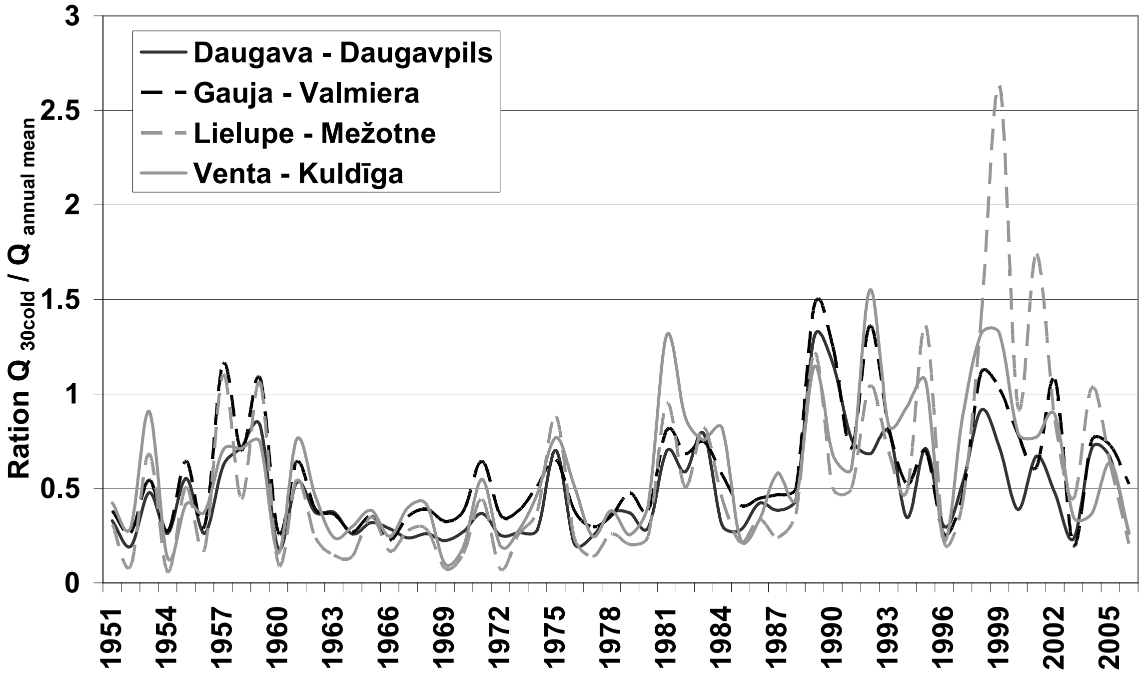

Figure 3. The trends of rivers' 30-day minimum discharge of could period discharge of the year from 1951 to 2006.

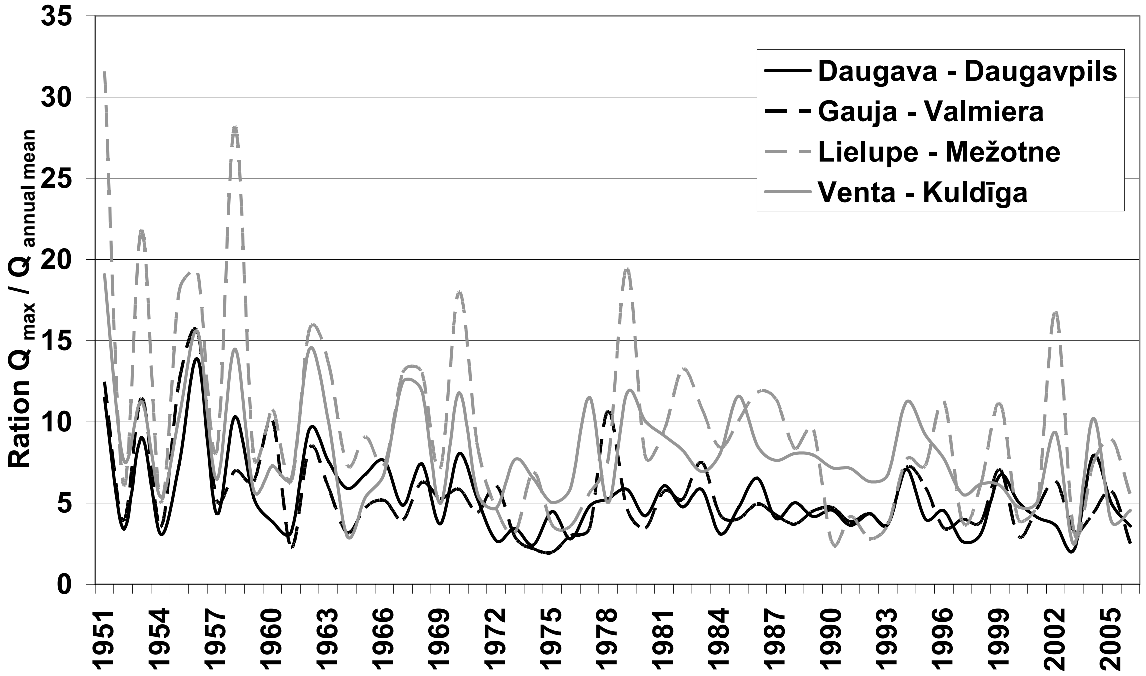

Figure 4. The trends of the maximal discharge of the year from 1951 to 2006.

CONCLUSIONS

The present results confirm the basic hypothesis that main tendency in runoff changes are following: decreased in spring flood and as opposite - increased in winter thaw-period for the study period 1951-2006. The analysis of obtained results shows regional differences in total annual river runoff distribution among, high and low flows by Western, Central, Northern and Easter parts of Latvia.

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INTEGRATED FRAMEWORK FOR ASSESSING UNCERTAINTY IN CATCHMENT-SCALE MODELLING OF CLIMATE CHANGE IMPACTS: APPLICATION TO PEAK FLOWS IN FOUR NORWEGIAN CATCHMENTS

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ABSTRACT

Assessment of catchment-scale hydrological impacts of climate change is often based on models driven by input timeseries derived from climate scenarios. Input data are obtained from one or several climate models under one or more emission scenarios and are adjusted using downscaling techniques to most closely reflect local conditions. Each step in this model chain from the selection of an emission scenario to the interpretation of a hydrological impact introduces uncertainty in the projected outcome. In this paper, we present a methodology for representing uncertainty in impact model outcomes (in this case, q_{95} in the period 2071-2100) based on the application of the HBV model using input data representing two emission scenarios (A2 and B2), two GCMs (Hadley and ECHAM), and two downscaling techniques (empirical adjustment and delta change methods). Multiple HBV model parameter sets (generated using PEST parameter estimation routines) are used such that hydrological parameter uncertainty is also incorporated in the outcomes. Four catchments (Flaksvatn, Viksvatn, Masi and Nybergsund), representing different seasonal flow patterns, are used to demonstrate the methodology for presenting uncertainty and to illustrate their differing sensitivities to climate model input data relative to HBV model parameter uncertainty.

INTRODUCTION

The potential impacts of climate change on runoff have significant implications for the optimisation of future hydropower production capacity, for the development of flood risk management plans, and for the assessment of reservoir dam safety throughout the Nordic region. Analyses of future changes in runoff are necessarily reliant on the application of a cascade of models, commencing with a general circulation model (GCM) run under a given emissions scenario (*e.g.* A2 or B2), dynamically downscaled to a regional level using, for example, the HIRHAM model. In order to derive daily data of a quality suitable for hydrological modelling, the climate scenario data is further downscaled to a local level, *e.g.* meteorological station sites. At each level of this model chain, one or more alternative models may be available, such that the end product of an impact analysis is actually a range of possible outcomes, rather than a 'one-number' estimate for a particular catchment. In addition to uncertainty derived from climate scenario data, different hydrological models, as well as different parameter sets for a given hydrological model, can also produce different outcomes. Consequently, this array of projected outcomes must be analysed and presented in a manner that conveys both the expected changes in runoff and the range of uncertainty underlying those estimates, if the results are to be of use in decision making and planning. In this paper, we evaluate potential climate change impacts on peak flows (q_{95}) in four Norwegian catchments, using the probabilitistic framework for impact analysis previously presented by Wilby and Harris (2006). Uncertainties introduced by choice of emission scenario, GCM, downscaling technique and HBV model parameter set are incorporated in the analysis and are presented relative to the magnitude of expected changes in runoff in each of the four catchments.

METHODOLOGY AND MODEL APPLICATION

The 'Nordic' version of the HBV model (Sælthun, 1996; based on Bergström, 1976; 1995) was applied to the four catchments illustrated in Figure 1.

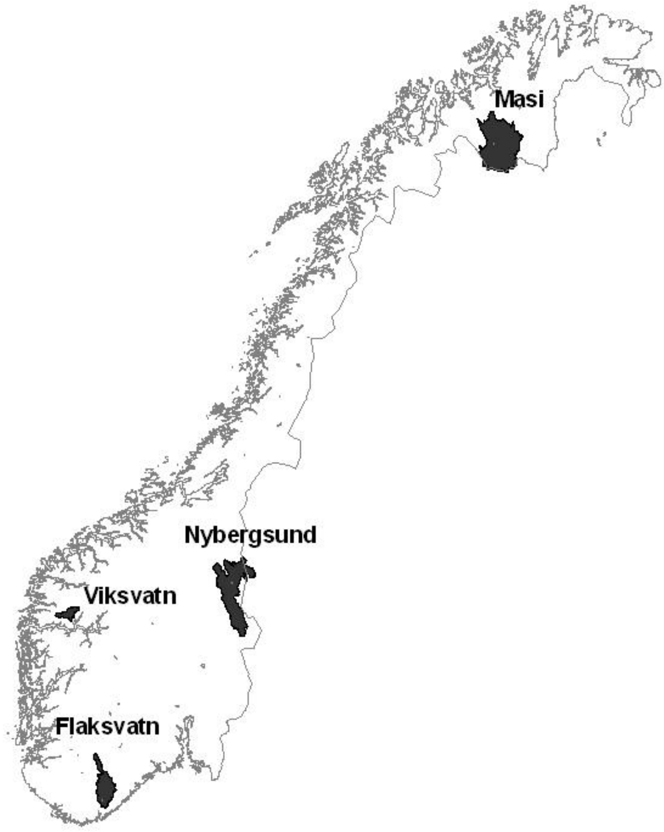

Figure 1. Location of catchments used in the analysis of projected changes in peak flows.

The four catchments represent differing flow regimes and projected climate change impacts on peak flows (Beldring, *et al*., 2006; in press). High flows in Masi, Nybergsund and Viksvatn occur principally in spring or early summer due to snowmelt, whereas Flaksvatn is also subjected to autumn high flows, in addition to spring snowmelt floods. Under climate projections for 2071-2100, earlier snowmelt and reduced snow storage will lead to a reduction in snowmelt floods in Nybergsund and Masi. In Viksvatn and Flaksvatn, an increase in both precipitation and the proportion of precipitation which falls as rain, will contribute to an increase in peakflows (Beldring, *et al*., in press).

The HBV model was calibrated and validated, relative to observed daily discharge, in each of the catchments for the control period (1961-1990), with 1970-1985 as the calibration period and the remaining years as validation periods. Observed precipitation and temperature data were derived from local station data, interpolated and corrected for altitude differences, using the GWB (gridded water balance) model (Beldring, *et al.,* 2003). PEST automatic parameter estimation routines (Doherty, 1998) were applied to generate 150 best-fit model parameter sets, based on the parameter ranges given in Table 1.

Table 1. HBV parameter ranges used in PEST optimisation.

HBV Parameter	Description	Range considered
BETA	Soil moisture parameter	1.0 – 4.0
CX	Degree day correction factor	1.0 – 5.0
FC	Field capacity – soil zone	50.0 – 500.0
KLZ	Recession constant – lower zone	0.001 – 0.1
KUZ1	Recession constant – upper zone 1	0.01 – 1.0
KUZ2	Recession constant – upper zone 2	0.1 – 1.0
PERC	Percolation – upper to lower zone	0.5 – 2.0
PGRD	Precipitation lapse rate	0.0 – 0.1
PKORR	Rainfall correction factor	0.8 – 3.0
SKORR	Snowfall correction factor	1.0 – 3.0
TS	Threshold temp. for snowmelt	-1.0 – 2.0
TX	Threshold temp. for rain/snow	-1.0 – 2.0
TTGD	Temp. lapse rate – Clear days	-1.0 – - 0.5
TVGD	Temp. lapse rate during precip.	- 0.7– - 0.3
UZ1	Threshold for quick runoff	10.0 – 100.0

In the methodology used for parameter optimisation, initial values for the parameters were randomly selected and the parameter set was tested in a trial calibration run. Only those parameter sets which gave Nash-Sutcliffe (N-S) values >0.25 were used in the subsequent PEST optimisation run. The final Nash-Sutcliffe values for the optimised parameter sets for model validation periods varied somewhat between catchments and over the range of the 150 parameter sets for each catchment (Flaksvatn, 0.76-0.79; Masi, 0.83-0.86; Nybergsund, 0.88-0.91; and Viksvatn, 0.85-0.87), but in all cases they represent very good model fits. The most sensitive parameters also varied between the catchments, with KLZ (lower zone recession constant) and PKORR (rainfall correction factor) being the most sensitive parameters in Flaksvatn and Viksvatn and PKORR and SKORR (snowfall correction factor) dominating the sensitivity in Masi and Nybergsund.

Following the calibration of the parameter sets for each catchment, precipitation and temperature series derived from climate scenarios were used as input to the HBV model. The analyses of climate change impacts presented here are based on the HadAM3H A2 and B2 and the ECHAM/OPYC3 B2 scenarios, dynamically downscaled in the RegClim project via the HIRHAM regional climate model (Bjørge *et al.*, 2000). These GCM models are referred to hereafter simply as the Hadley and the ECHAM models. Temperature and precipitation series from the regional climate models for each of the scenarios available (Hadley A2 and B2, ECHAM B2) were downscaled to meteorological station sites using two approaches: the delta change method (Reynard *et al.*, 2001) and an empirical adjustment technique (Engen-Skaugen, 2007). Accordingly, six temperature and precipitation series for the period 2071-2100, representing two GCMs, two emission scenarios and two methods for local downscaling were used. The empirical adjustment technique was also used to downscale temperature and precipitation series for the control period (1961-1990) scenarios for the Hadley and ECHAM models, for comparison with the observed data for this period. The HBV model was run for these input timeseries in each catchment for each of the 150 calibrated parameter sets, and cumulative distribution functions were constructed from the daily discharge series from each model run. From these distribution functions, combined cumulative distribution functions were constructed for a given percentile level (in this case, the 95th percentile), such that the full range of model results at that particular flow level is represented. The term q_{95} is thus used here refer to the probability of non-exceedance, such that 95% of the daily flows are equal to or less than the value given.

RESULTS

The cumulative distribution functions for the 95th percentile runoff (mm/day), q_{95}, for the control period runs (1961-1990) are illustrated in Figure 2 for the four catchments and show the range of q_{95} values for all 150 model runs based on the input dataset indicated. In three of the catchments, Flaksvatn, Nybergsund and Viksvatn, the input timeseries from the ECHAM model more closely reproduces results obtained from the observed P, T values than does the Hadley model. In Masi, the Hadley model output provides a closer correspondence. The range of values obtained for a particular input dataseries represents the uncertainty introduced by the model parameter sets (*i.e.* hydrological model uncertainty) whereas the differences between the three series illustrate the uncertainty introduced by the use of a climate model. If one considers the range of values between 5 and 95% for each function and compares this with the differences in the 50% value for the functions, the relative significance of the two contributions to uncertainty can be assessed for each catchment. Accordingly, in Masi and Nybergsund, HBV model parameter uncertainty is secondary to that associated with the use of climate model data, whereas in Flaksvatn they are of a similar scale and in Viksvatn HBV model parameter uncertainty is more significant.

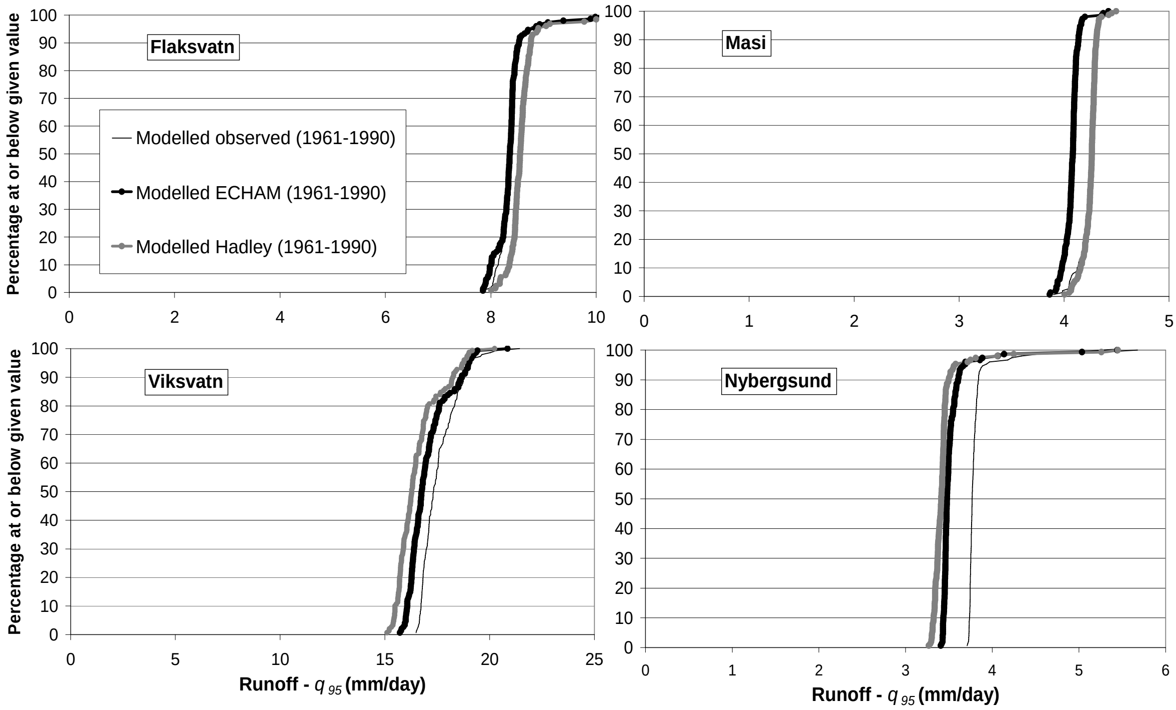

Figure 2. Cumulative distribution functions for q_{95} values for all model parameter sets for the input P, T dataseries (Hadley – thick grey line; ECHAM – thick black line; modelled observed – thin line). The 'modelled observed' refers to models run with the observed P, T dataseries for the control period. In Flaksvatn, the modelled observed and ECHAM series coincide and in Masi, the modelled observed and Hadley series coincide, such that the modelled observed series is not visible in the figure. The term q_{95} refers here to the probability of non-exceedance, such that 95% of the daily flows are equal to or less than the value given.

Cumulative distribution functions for the 2071-2100 model runs are illustrated and compared with the control period (1961-1990) runs in Figure 3. The modelled observed values are also shown for reference. The 'all scenarios' curves represent the combined distribution functions for the scenarios available for that time period (two for the control period, each illustrated in Figure 2, and six for the period 2071-2100). All of the distribution functions were scaled by the median value for the modelled observed (*i.e.* for the control period 1961-1990), so that the results are given as a percentage difference from that reference. Accordingly, the 'difference' between the 'all scenarios' curve for 1961-1990 and the 'modelled observed' reflect the discrepancies between the ECHAM and Hadley climate model output and the output based on the observed P, T, timeseries illustrated in Figure 2.

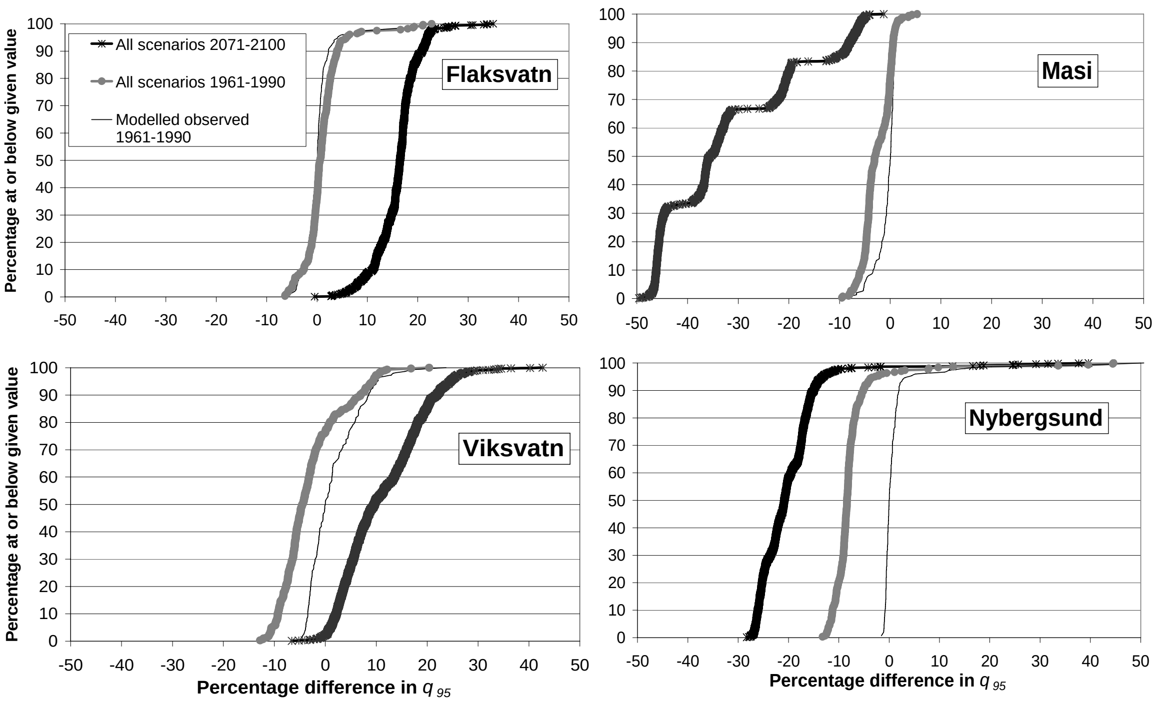

Figure 3. Cumulative distribution functions for q_{95} values for the combined distributions for the scenarios for the control period (1961-1990; thick grey line) and the period 2071-2100 (thick black line). Modelled values based on observed P and T for the period 1961-1990 (thin black line) are the same as those shown in Figure 2 are also shown here for reference.

If we consider only the median value for each distribution shown in Figure 3, then the results clearly indicate an increase in q_{95} in Flaksvatn and Viksvatn and a decrease in Masi and Nybergsund between the control period and 2071-2100, in agreement with previous analyses (Beldring *et al*., 2006). The range of modelled values, however, varies significantly between catchment, as does the underlying uncertainty they represent. The results for Flaksvatn indicate an increase in q_{95} of ~ 8 to 22% relative to the reference value (median modelled observed) and based on the values between 5 and 95% of the distribution. This range of values includes uncertainty due to choice of HBV model parameter set and to choice of climate scenario (*i.e.* A2 *vs*. B2, ECHAM *vs*. Hadley, and delta change *vs*. empirical adjustment downscaling) based on the six 2071-2100 scenarios used here. In the case of Flaksvatn, there is little overlap between the ranges for the control period and the 2071-2100 scenarios, such that the outcomes in each timeslice are distinct. This is also the case for Masi and Nybergsund (for the range between 5 to 95%), and the results indicate a decrease of 8 to 47% and of 16 to 28% relative to the reference value in the two catchments, respectively. The wide range of outcomes for the Masi catchment reflect the highly variable results derived from the climate scenarios for this catchment. The stepped nature of the combined function shown represents largely non-overlapping distributions for the range of values for each modelled scenario. Nybergsund exhibits a much narrower range of outcomes (excepting the values above 95%) and the smoother distribution function reflects more overlap between the underlying models. In this catchment, however, the control period runs have q_{95} values that are ~ 8 to 11% less than the reference value, such that the projected decrease in q_{95} based on the timeslices represented by the climate scenarios is somewhat less (i.e. of the order of 10-20%). The results for the Viksvatn catchment are the most ambiguous in that the range of projected increases in q_{95} range is 2 to 23%, overlapping slightly with the control model runs (-10 to +7% of the median modelled observed value).

DISCUSSION

Figures 2 and 3 represent a relatively simple methodology for presenting the results of a climate impact analysis derived from multiple climate scenarios, which also incorporates hydrological model parameter uncertainty. They also illustrate significant variability in the range of outcomes and in the contribution of climate scenario input data *vs*. HBV parameter model parameters to this range in the four catchments considered. The approach used here is based on an equal weighting (*i.e.* unconditional probability) for the six 2071-2100 climate scenario datasets available for this work and is used to display the range of outcomes. A more detailed approach can be adopted for individual catchments by, for example, weighting the Hadley *vs*. the ECHAM results relative to their performance in the control period (Figure 2) and using these weightings in constructing the 2071-2100 distributions (Figure 3). Additionally, the contributions of emission scenario *vs*. GCM *vs*. downscaling technique to uncertainty in the 2071-2100 projections varies between catchments (thus the differences between the shape of the curves for Flaksvatn and Masi) and this will also be considered in more detail in further work.

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OVERLAND FLOW INDUCED BY SNOWMELT, EFFECTS OF CLIMATE CHANGE

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ABSTRACT

The transport infrastructure is a vital part of the society, with high capital investments. Design of this system is therefore very important. Climate changes will increase the frequency of extreme precipitation events, floods and snowmelt periods experienced by the infrastructure. Increased frequency of floods is expected to cause more closed roads because of insufficient maintained drainage systems. Increased ground frost and ice formation on ground surface cause large increases in surface runoff during snowmelt. Recently, in Norway the ClimRunoff project has started focusing on quantifying discharge of catchment areas draining towards roads. A model will be adapted that can evaluate the run-off situations under spring situation (i.e. overland flow due to snowmelt and partially frozen soils). The model is tested on a well-defined catchment under winter situation and will be adapted towards spring situation. This means that snowmelt routines and infiltration in frozen soils will be added to the existing LISEM model. Results of the measurements and model calculations, calibration and validation shows that he model is potentially capable in calculating discharge from small agricultural catchments in Norway, even for different climate scenario's.

INTRODUCTION

The transport infrastructure is a vital part of the society, with high capital investments, proper management and design of this system is therefore very important. Climate changes will increase the frequency of extreme precipitation events, floods and snow melt periods experienced by the infrastructure. Expected climate changes in Norway according to the Norwegian research project RegClim (e.g. Benestad and Førland, 2004) include:

- Increased temperatures, largest increases during autumn and winter

- Increased precipitation, largest increases in autumn and winter on the west coast
- Increased frequency of extreme events such as intensive rain episodes, and freeze/thaw incidents, and change in the timing of freezing and snowfall
- Less effect on summer conditions

According to initial analysis by the Norwegian transport sector (Nasjonal Transportplan 2006- 2015, Anonymous, 2002) these changes will affect, road maintenance, emergency planning, design of new roads and infrastructure. Increased frequency of floods is expected to cause more closed roads because of insufficient and badly maintained drainage systems. More erosion on areas next to the roads is expected because of higher groundwater levels, which may cause instability of road fundaments, and also clog existing drainage systems. More ground frost is expected to affect the road quality and lifetime. Increased ground frost and ice formation on ground surface may also cause large increases in surface runoff during snowmelt (Johnsson and Lundin, 1991).

Changes in winter climatic conditions will have large effects on hydrology in general and subsequently influence the local hydrology near roads, runways and buildings causing changes in:

- Water and energy balance, which determines when liquid water is available, hence it strongly affects the melt water management at a particular location
- Pollution risk, as the amounts and distribution of water and diffuse pollutants on the surface determine the potential for retention (degradation, adsorption and uptake by vegetation) of pollutants before melt water reaches groundwater or surface waters

Physical processes in the soil occurring at cm^3 and m^3 scale near the surface, are important controls of the flow of meltwater and can also affect the hydrology at a catchment scale (e.g., the response time). According to e.g. Baker and Spaans (1997); French and van der Zee, (1999); Johnsson and Lundin (1991); French and Binely (2004), infiltration during snowmelt often occurs as focused recharge in local depressions on the surface. This may cause higher velocities through the unsaturated zone than during evenly distributed infiltration on the surface, hence causing less than optimal conditions for degradation of pollutants. The redistribution of meltwater can also cause extreme runoff in areas where there is normally high infiltration capacity. Hence it is important to take the dynamics of infiltration capacity into account in estimates of produced runoff from catchment areas draining towards roads. Damages for several million Norwegian kroner, have been caused by extreme storm events.

In 2007, the Bioforsk started the CLIMRUNOFF project. The project is carried out in close cooperation with Luleå University of Technology, Sweden, the University of Minnesota, USA. The project is financed by the Norwegian Research Council. Main objectives of the project are to:

- Classify regions of Norway with respect to the vulnerability of road infrastructures to extreme run-off situations, with special focus on winter and spring conditions
- Improve modeling tools for improved prediction of run-off situations near roads
- Improve construction and dimensioning guidelines for road construction, maintenance and drainage considering possible consequences of climate change

This paper describes the work carried out so far with emphasis on the study area, model choice, initial model results and limitations.

METHODS

Modelling

For modelling discharge from small catchments, a number of models are available. A quick investigation has been done for a number of models. The models considered are: WATER, LISEM, SUTRA, WEPP, SWAT and COUP. None of the described models meet all defined criteria. The WATER model is capable of calculating infiltration using several methods, but the spatial scale is limited. LISEM calculates infiltration and soil loss and nutrient losses on catchment scale. The timescale is event based. However, LISEM is not capable in calculating sub-surface flow and snowmelt input. SUTRA is not able to calculate overland flow and sediment transport. WEPP is able to calculate effects of frozen soils and snowmelt. However, catchment scale is generalized by uniform hill slopes, leaving no option for field heterogeneity. The same goes for the SWAT model, where HRU's are defined, considered homogeneous. Besides this, the SWAT model is not capable in calculating rainfall events, but generalizes results on a daily basis. In the SWAT model, infiltration is based on the curve number method or Green-Ampt equation. Overland flow in the WEPP model is calculated as Hortonian flow, and sediment transport based on USLE-like empirical assumptions. A study on Norwegian catchments (Grønsten and Lundekvam, 2006) showed that the WEPP model did not produce satisfactory estimates of surface runoff and soil loss.

Given this very brief analysis and given the experience gathered with the models, first focus of the ClimRunoff project is on using the LISEM model. Assumptions are:

- flooding is event based
- no effect of subsurface flow on total discharge during an event

Consequences are that the model should be extended with (i) a snowmelt routine (ii) infiltration in freezing soils (iii) effects of drainage system should be quantified.

The LISEM model uses the Richards' equation to calculate the infiltration. A soil profile of 1 m depth is assumed, with an estimated 25 cm thick frozen layer forming the upper layer of the soil profile. The majority of the input parameters are gathered from Deelstra et al. (2005).

Study area

The model performance is initially being tested on an agricultural catchment close to Ås: the Skuterud area. The catchment is 4.5 km^2 in size, dominated by agricultural land. Part of the catchment is urban area and forest. The outlet of the area crosses the E18 near Holstad. The Skuterud area is part of the JOVA program, which for more than 10 years monitors discharge, sediment load, water quality and agricultural practices in several catchment in Norway (Bechmann et al., 1999).

Calibration

An event on January 10 2008 is used to calculate the hydrograph. This event took place after a long period of frost, resulting is a frozen top layer of the soil. Calibration of model results have been performed on measured peak discharge optimizing the saturated conductivity value of the frozen layer. The calibrated saturated conductivity value has been used to calculate the rain event from 13-16 January 2008 for validation. A view of the overland flow processes during the event on January 16, 2008 is shown in Figure 1.

Figure 1. Overland flow in the Skuterud area during the rain event on January 16, 2008.

Scenario analyses

Scenario analyses predicts the effects of different situation in the area. For this project, the initial scenario's are focusing on possible climate change scenario's. These scenario's will be defined on down-scaling climate change models to the study area. As an initial step to this, we have increased the rainfall intensity with 10 and 20% respectively to estimate the effect on discharge. Another important scenario is land use. By incorporating measures to reduce soil erosion and water losses in the model input, their effect can be quantified and analyzed.

RESULTS

The event of January 10 had a total rainfall of 13 mm with a maximum rainfall intensity of about 8 mm/h. The event from 13-18 January produced 80.6 mm total rainfall, and a maximum intensity of about 15 mm/h. Results of calculations of both events are shown in Figures 2 and 3. Fig. 2 shows the calibration, where 1 value for the saturated conductivity is optimized. After this, the model was adapted with a baseflow calculation. TFig. 3 shows the validation of the calibrated dataset.

To estimate effects of climate change scenario's, the rainfall intensity was increased with 10 and 20% respectively. Results of this calculation are show in Fig. 4.

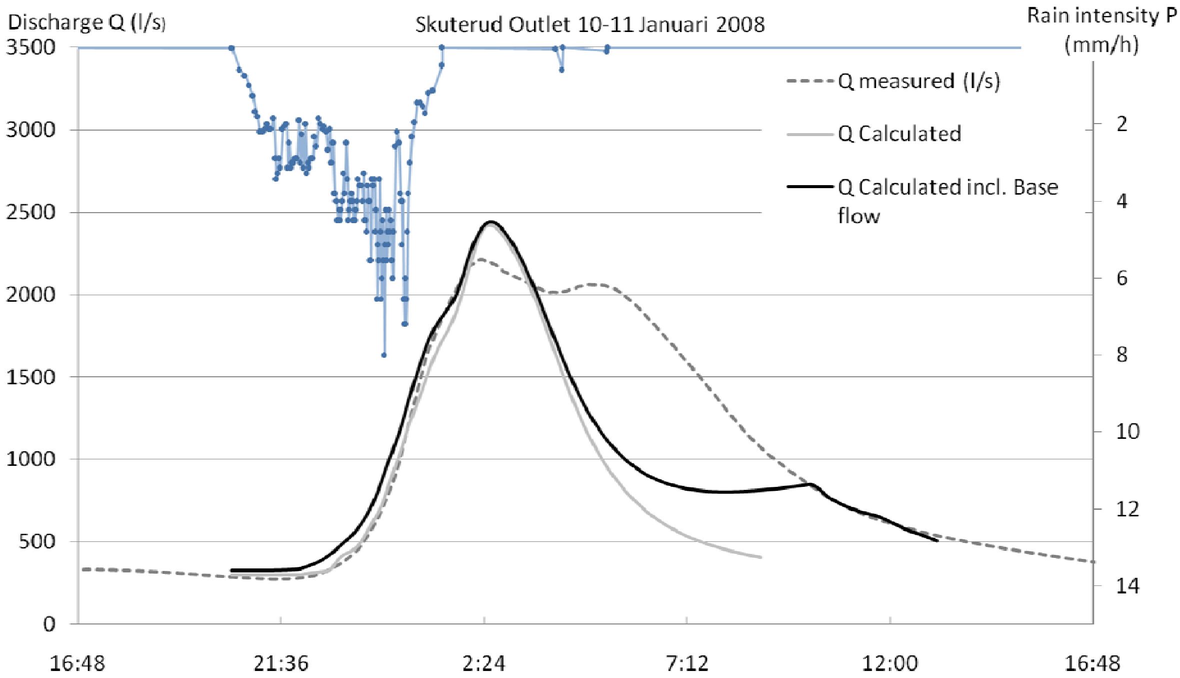

Figure 2. Measured and calculated hydrograph for the Skuterud area for a rain event on January 10, 2008. Black line is the measured hydrograph (5 min interval), calculations are done using the LISEM model for three different saturated conductivity values for the frozen top layer.

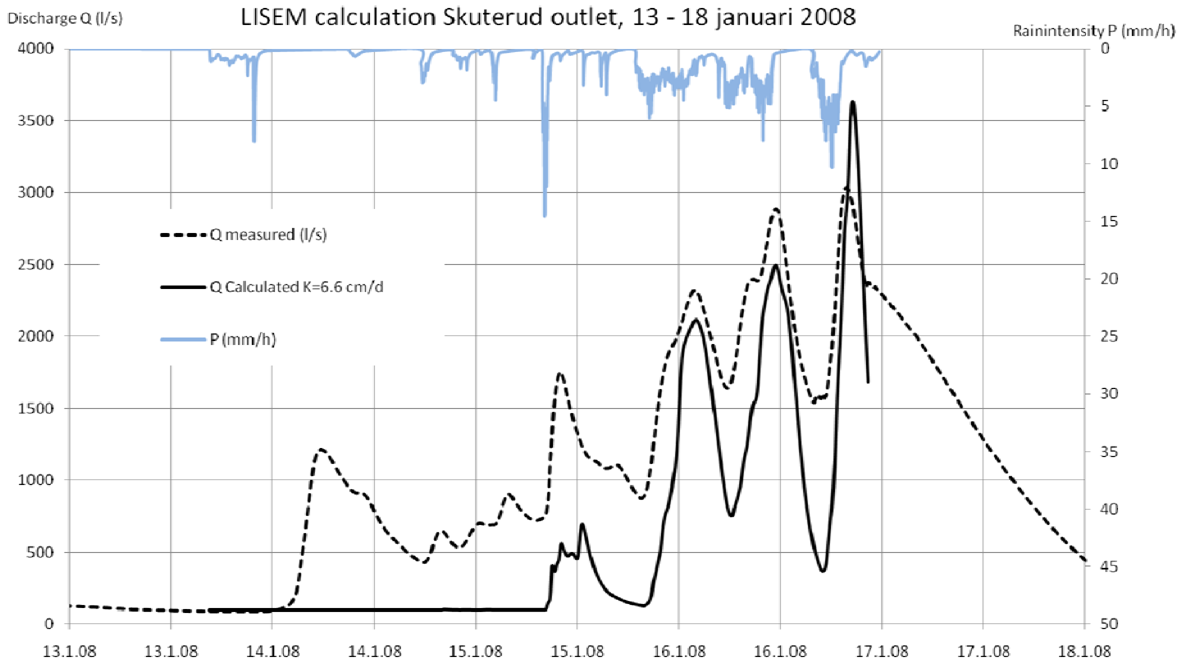

Figure 3. Measured and calculated hydrograph for the Skuterud area for a rain event on January 13-16, 2008. Blue line is the measured hydrograph (5 min interval), calculations are done using the LISEM model.

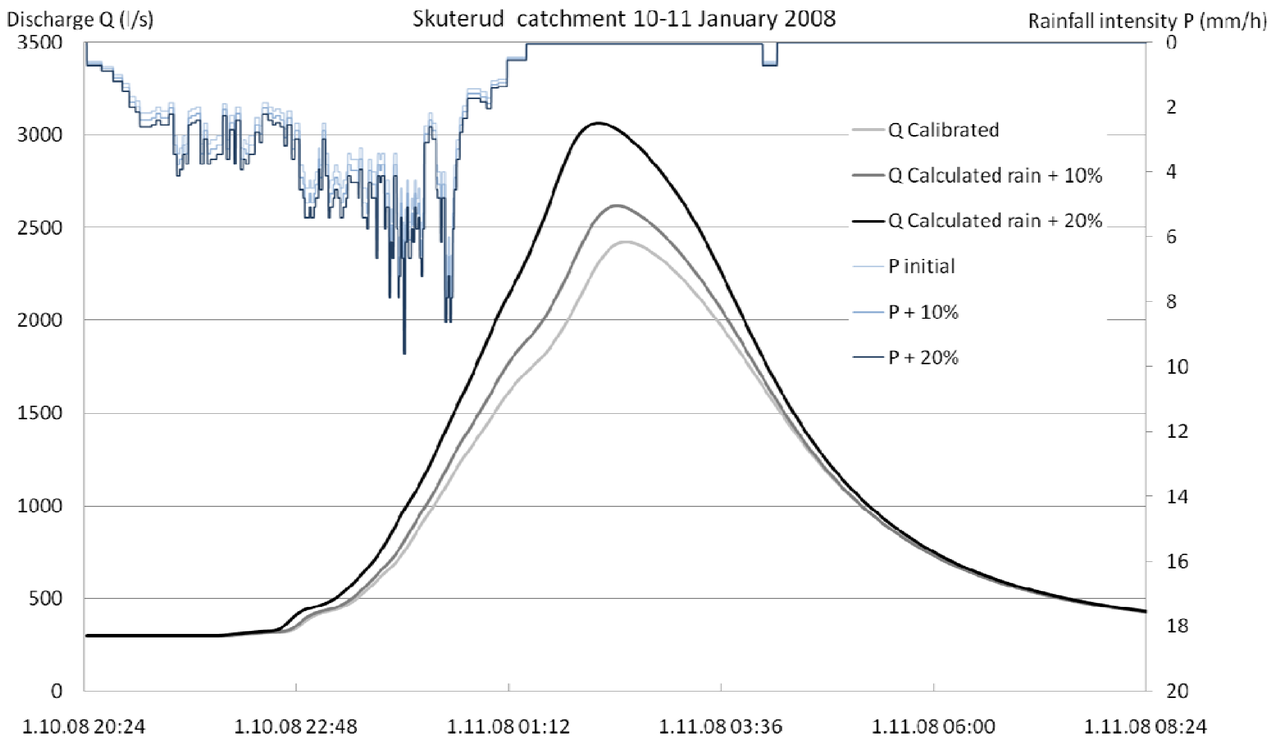

Figure 4. Calculated hydrographs for the Skuterud area for a rain event on January 10, 2008 and two rianscenarios, one eith 20% increase in rainintesity and one with 20& increase.Bleu lines are the rainevents, orange ines the calculated discharge. Th elight onage line is the calibrated event.

CONCLUSION AND DISCUSSION

Figure 2 shows a good estimate of the peak discharge. It can also be conluded that the model is very sensitive for the saturated conductivity, as is shown before by e.g. Stolte et al. (2003). The total amount off discharge is less accurate predicted by the model. This might be explained by the snowmelt, since the area was still partly covered by snow during the rain event. This snowmelt amount is not incorporated in the input data for the model. The addition of baseflow calculation improved modelling results considerable. From Fig 3, it shows that a calibrated parameter set is able to predict other events in term of peak discharge and peak time accurately. But, the first measured discharge caused by the event is not calculated by the model. This might be caused by the sub-surface drainage system on the agricultural fields. This sub-surface drainage process is not incorporated in the model, and might be the cause of the first measured discharge peak. Fig. 4 shows an increase of peak-discharge amount with increase of rainfall intensity. This increase however is not linearly. It also influences the peak time of the discharge. Both increases are of influence on road constructions and maintenance.

Conclusion from this initial survey on modelling results is that the LISEM model is potentially capable in calculating discharge from small agricultural catchments in Norway. A number of adaptations to the model have to be done to fine-tune the model for cold climate regions. Major effort will be on the quantification of snow melt; infiltration capacity of frozen layers; quantification of water through drainage systems. These modifications will be carried out during the ClimRunoff project.

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STUDY OF HYDROLOGICAL PROCESSES FOR DEVELOPMENT OF MATHEMATICAL MODEL METQ

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ABSTRACT

In this study, the last version METQ2007BDOPT of the conceptual rainfall-runoff METQ model simulating the daily runoff was applied. The model structure and parameters were basically the same as in the METQ98 with some additional improvement and semi-automatic calibration performance. The model METQ2007BDOPT was calibrated for five river basins of different size (the Vienziemīte, the Salaca and the outreach to the Bērze, the Imula and the Iecava) and the Lake Burtnieks watershed as the key-study drainage area of the model. For the Lake Burtnieks watershed the calibration results from 1994 to 1999 are the following: the Nach-Sutcliffe efficiency R^2 varies from 0.90 to 0.58 and correlation coefficient r - from 0.95 to 0.81. The preliminary results of calibration showed a good coincidence between the measured and simulated daily discharges. The R^2 varies from 0.86 to 0.50 and r from 0.91 to 0.71 for the five river basins with a calibration period from 1961 to 1990. However, we have obtained lower statistical criterions of the model validation to compare with calibration results in all study cases.

INTRODUCTION

In fact, for the purposes of water resource management, flood forecasting, water quality or future climate change assessment and many others, the availability of hydrological measurement data is limited in both temporal and spatial aspects. One possible method is the use of conceptual rainfall-runoff models which are widely used tools in hydrology (Seibert, 1999; Uhlenbrook et al., 1999; Beven, 2001; Zīverts and Apsīte, 2001) and usually simple and relatively easy to apply (Bergström, 1991; Merz and Blöschl, 2004).

In Latvia, during the last twenty years, several versions of mathematical models of hydrological processes have been developed – METUL (Krams and Ziverts, 1993), METQ96 (Ziverts and Jauja, 1996), METQ98 (Ziverts and Jauja, 1999), METQ2005 and METQ2006. In this paper the modelling results of the latest version METQ2007BDOPT are presented. The METQ is conceptual rainfall-runoff model of catchment hydrology, originally developed using Latvian catchments. The model is successfully applied to small and relatively large catchments, the Brook Vienziemīte (A=5.92 km^2) and the River Daugava (A=81,000 km^2 at the Pļavinas HPP) respectively. Furthermore, the METQ model has been used for different hydrological tasks, i.g., to evaluate the model performance before and after drainage construction and to estimate the probable maximum flood (Ziverts and Jauja, 1999), to study eutrophication and hydrotechnical problems of lakes, including climate change effects (Bilaletdin et al., 2004; Ziverts and Apsite, 2005), and to attempt model parameter sets for ungauged catchments from measurable variables and to simulate nutrients runoff in typical agricultural river basins (Jansons et al., 2002; Apsite et al., 2005).

The aim of this study was to apply a conceptual rainfall-runoff METQ model, the latest version METQ2007BDOPT, to five different size river basins and the Lake Burtnieks watershed as a key-study drainage basin for simulation of daily runoff and water level.

MATERIALS AND METHODS

The conceptual rainfall-runoff METQ model

The METQ model is a conceptual rainfall-runoff model of catchment hydrology which simulates daily discharge and evapotranspiration, using as input variables: daily air temperature, precipitation and vapour pressure deficit observations. The model consists of different routines representing snow accumulation and ablation, water balance in root zone, water balance in the groundwater and capillary water zone and runoff routing. Runoff routing can be simulated by simple hydrological methods, such as modifications of the unit hydrograph approach. However, if there is a lake in the river basin which considerably influences the hydrological regime of the river, then there is a need for hydraulic runoff routing. Furthermore description of model (METQ96 and METQ98) can be found elsewhere (Ziverts and Jauja, 1996; Ziverts and Jauja, 1999). In general, the structure and simulation of hydrological processes by the METQ98 model are similar to the HBV (Bergström, 1976; Bergström, 1992) model developed in Sweden. The main difference between the METQ98 and HBV models is that the degree-day ratio in METQ does not have a constant value, but it has a temporal difference depending on the daily potential insolation of each particular day (Ziverts and Jauja, 1999).

The latest version METQ2007BDOPT and its application

The METQ2007BDOPT model parameters are basically the same as for the METQ98 (Ziverts and Jauja, 1999). However, METQ2007BDOPT has one additional Beta parameter, providing twenty three parameters in total. Snow accumulation and melting is characterised by the following parameters: T_1 –daily mean temperature ^{0}C, at which snow accumulation starts; T_2 – daily mean temperature at which snow melting starts; CMELT – degree-day ratio and characterises intensity of snow melting; AMELT – conversation factor which increases degree-day ratio on the daily potential insolation of each particular day; KS – evaporation coefficient from snow; WHC and CFR characterise snow accumulation and melting processes. The water balance from root zone is characterised by: WMAX – threshold value of water storage in root zone (mm); KU and KL – coefficients characterising intensity of evaporation from root zone; RCHR (mm/d), RCHRZ (mm/d), RCHR2 (mm/d), RCHR2Z (mm/d) and ROBK – characterise the infiltration capacity of soil. The water balance of groundwater storage and runoff is characterised by the following parameters: ALFA – fillabale porosity of the aquifer; ZCAP – height of capillary rise (cm); DZ – depth of upper level drain from the surface (cm); A2 and Beta characterize daily subsurface runoff Q_2 of upper level „drain"; PZ characterises depth of the lower level "drain" (cm); A3 – daily runoff Q_3 of the lower level "drain"; DPERC is intensity of deep percolation to the aquifers (mm/d). Most of the parameters in the model are physically based and the rest of parameters usually can be estimated by the calibration. The METQ2007BDOPT has semi automatic calibration performance for the following parameters – A2, DZ, A3, PZ, CMELT, AMELT, DPERC, Beta, RCHR, RCHR2, RCHRZ, and RCHRZ2.

To consider the runoff heterogeneity in the runoff processes, the studied catchments were divided into hydrological response units (HRU) characterised by a relative homogeneity with respect to the most important parameters, including slope, vegetation and soil characteristics. Catchment area was divided into six HRUs: agricultural lowlands, hilly agricultural lands, forests, swamps, lakes and sandy lowlands. The last one as an important additional geomorphologic HRU was introduced in the METQ2005, the METQ2006 and the METQ2007BDOPT versions what improved the modelling results of runoff in likely sandy catchments. The water balance and runoff of each HRU has been simulated in three storages: snow (water content in snow cover), soil moisture (water in the root zone) and groundwater. The total runoff from each HRU consists of three runoff components: Q_1 – surface runoff, Q_2 – subsurface runoff (runoff from the groundwater upper zone) and Q_3 – base flow (runoff from the groundwater lower zone).

Daily meteorological data were used as input data for the METQ2007BDOPT model. Measurements of air temperature, precipitation and vapor pressure deficit at eleven meteorological stations were used for preparing the climatic data series. For the model calibration and validation of daily river discharge and water level of the lake (in this study – the Lake Burtnieks) of nine hydrological gauging stations were applied. A statistical criterion R^2 (Nash and Sutcliffe 1970), a correlation coefficient r, mean values and graphical representation were used in the analyses of model calibration results.

The Lake Burtnieks watershed was selected as a key-study for the test of the latest version of METQ2007BDOPT. The selected calibration period was from 1994 to 1999 and validation period – from 2000 to 2005**.** The Lake Burtnieks is the fourth largest lake in Latvia. Its total drainage is 2215 km^2 and it occupies 62 % of the River Salaca catchment; the surface area of the lake is 40.06 km^2. The lake watershed was divided into four sub-basins (the Rūja, the Seda, the Briede and the Burtnieks (small rivers entering the lake) and one additional sub-basin between the outlet of the lake and the runoff gauging station Mazsalaca at the River Salaca. The modelling of the Lake Burtnieks watershed sub-basins was done using the METQ2007BDOPT version by testing the following feasibility studies: (1) calculation of daily average discharge and water level in the Lake Burtnieks based on daily average temperature and precipitation data from the nearest meteorological stations; calculation of evapotranspiration from the drainage basin by using climatic parameters of the corresponding calendar days; (2) optimization of parameters (fitting to larger sub-basin) was performed using the observed data from three runoff gauges: Rūja – Vilnīši A=729 km^2, Seda – Oleri A= 431 km^2, Briede – Dravnieki A=369 km^2; (3) runoff routing of the Lake Burtnieks simulated on the bases of common hydraulic methods of the reservoirs and open channels: the relation between water level and volume was estimated from geometrical dimensions of the lake, and discharge rating curve Q=f(H) at the outlet of the Lake Burtnieks for the River Salaca was calculated as for a parabolic river bed with a parabolic parameter, defined by the simulated average daily discharge and the given water surface slope. There was a lack of channel measurements at the outlet of the lake and the Q=f(H) was obtained on the bases of typical parameters for river channels in Latvia (Golubovskis, 1993).

In this study additionally, the conceptual METQ2007BDOPT model was calibrated and validated to the following different size river basins at the gauging stations: the Salaca (A=3220 km^2), the Imula (A=232 km^2), the Vienziemīte (A=5.92 km^2), the upper reaches of the Bērze (A=904 km^2) and the Iecava (A=566 km^2). The calibration period was selected from 1961 to 1990 (30-years as the control climate) with an aim to simulate the scenario climate from 2071 to 2100 in the future, and validation period – next ten years from 1991 to 2000.

RESULTS AND DISCUSSION

In the present study the mathematical model of the Lake Burtnieks watershed was calibrated in two ways – (1) three river sub-basins of the Lake Burtnieks watershed were calibrated using following river gauging stations: Briede-Dravnieki, Rūja-Vilnīši and Seda-Oleri, and (2) the whole watershed of the Lake Burtnieks, calibrated for the Salaca-Mazsalaca river gauging station and the Burtnieks-Burtnieki lake gauging station.

In the first case, the results of model calibration for a 6-year period showed sufficient or even good coincidence between the observed and simulated daily discharges (Table 1). Correlation coefficient r is 0.90 and efficiency criterion R^2 = 0.78 for the River Briede at Dravnieki, r = 0.87 and R^2 = 0.66 for the River Seda at Oleri. A little weaker coincidence between observed and simulated hydrographs was found for the River Rūja at Vilnīši where r is 0.812 and R^2 = 0.60.

In the second case, the model calibration shows sufficient fluctuations between simulated and observed water levels of the Lake Burtnieks for a period from 1994 to 1999: the Nach-Sutcliffe efficiency R^2 is 0.58 and the correlation coefficient r = 0.83 (Fig. 4). However, a very good coincidence between the observed and simulated daily discharges was obtained for the River Salaca basin at Mazsalaca: the efficiency criterion R^2 is 0.90 and correlation coefficient r = 0.95 (Fig. 1).

Table 1. The obtained statistical criterions for the Lake Burtnieks watershed

Gauging station	Calibration period (1994-1999)		Validation period (2000-2005)	
	R^2	r	R^2	r
Briede - Dravnieki	0.78	0.90	0.69	0.87
Rūja - Vilnīši	0.60	0.81	0.52	0.75
Seda - Oleri	0.66	0.87	0.58	0.82
Salaca - Mazsalaca	0.90	0.95	0.81	0.93
Burtnieks – Burtnieks [1)]	0.58	0.83	0.46	0.83

[1)] – close since 2004

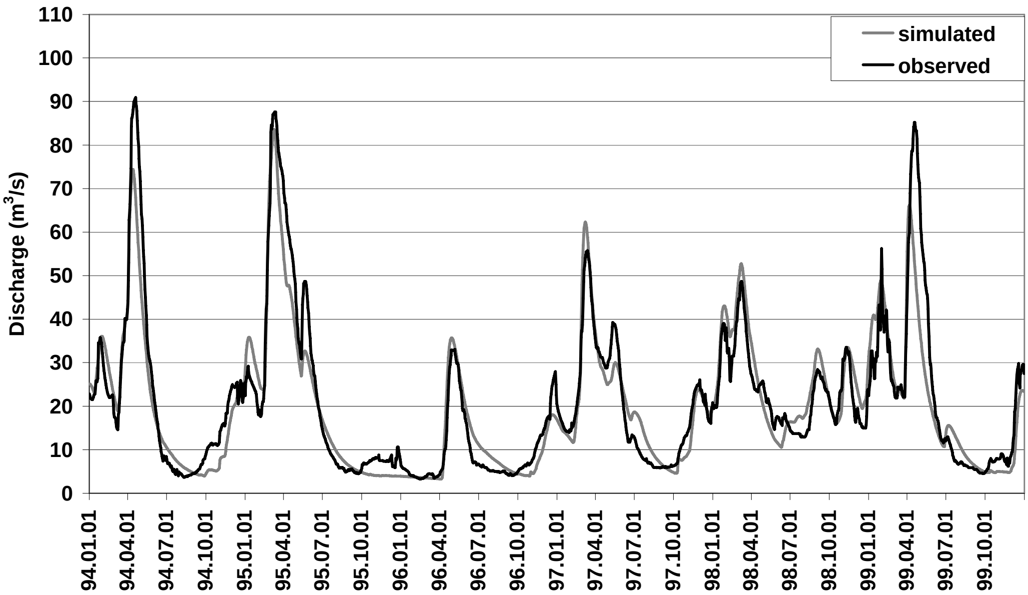

Figure 1. Observed and simulated daily discharge at river gauge Salaca-Mazsalaca of model calibration.

Table 1 shows the used values of model parameters for the river sub-basins of the whole Lake Burtnieks watershed, i.e. in the second case of the METQ2007BDOPT model calibration. The numerical values of the model optimized parameters for each sub-basin reflect some geomorphologic conditions of studied drainage area. The studied river sub-basins could be divided into two groups: (I) the Rūja and the Seda, and (II) the Briede, the Burtnieks and the Mazsalaca. The river basins of Rūja and Seda are located in the Burtnieks Plain and are characterised by a lower hypsometry, sandy and moraine areas, a high percent of bogs and flood plain in lower reaches. The River Briede basin is characterised by the moraine hilly topography and bog areas. The geomorphologic conditions of the sub-basins Burtnieks and Mazsalaca are more similar to the River Briede basin. However, the regionalization of the METQ2007BDOPT model parameters for river basins in Latvia could be another interesting issue for further research.

Table 2. Optimal parameter values of the METQ2007BDOPT model for the River sub-basins of the Lake Burtnieks watershed.

Parameters and unit[1]	The name of sub-basin				
	Briede	Seda	Rūja	Burtnieks	Mazsalaca
WMAX,mm	62	64	63	62	62
ALFA	0.163	0.170	0.150	0.163	0.163
ZCAP,cm	143	140	140	143	143
A2	0.00063	0.00058	0.00060	0.00063	0.00063
DZ,cm	84	81	72	84	84
A3	0.00075	0.00074	0.00075	0.00075	0.00075
PZ,cm	260	240	260	260	260
BETA	2.1	2.1	2.1	2.1	2.1
KU	0.58	0.59	0.63	0.58	0.58
KL	0.20	0.23	0.23	0.22	0.22
KS	0.05	0.05	0.05	0.05	0.05
CMELT	2.5	2.5	2.5	2.5	2.5
AMELTK	0.05	0.08	0.08	0.05	0.05
T1,^{0}C	0.5	0.5	0.5	0.5	0.5
T2,^{0}C	-0.1	-0.1	-0.1	-0.1	-0.1
RCHR,mm/d	18	16	16	18	18
RCHRZ,mm/d	25	25	25	25	15
RCHR2,mm/d	26	25	25	26	16
RCHR2Z,mm/d	20	20	20	20	20
ROBK	1.5	1.5	1.5	1.5	1.5
WHC	0.1	0.1	0.1	0.1	0.1
CFR	1.2	1.2	1.2	1.2	1.2
DPERC,mm/d	0.0	0.0	0.0	0.0	0.0

[1] - see the decryption of parameter abbreviation in Material and methods

The results of calibration showed good coincidence between the measured and simulated daily discharges. The Nach-Sutcliffe efficiency R^2 varies from 0.90 to 0.58 and correlation coefficient r from 0.95 to 0.81. The highest R^2 and r was acquired for the River Salaca at Mazsalaca. The lowest statistical criterions were found for the River Rūja and the Lake Burtnieks. The validation of model was done for the next 5-years period from 2000 to 2005, except the lake gauging stations Burtnieks-Burtnieks which was closed since 2004. We obtained lower statistical criterions for all gauging stations in this study (Table 1).

One of the main sources of difference between the simulated and observed runoff values in model calibration is the quality of precipitation input data and location of the available meteorological stations to characterise the spatial and temporal distribution of precipitation in the studied drainage area. For instance, the meteorological station Rūjiena locates in the middle part of the River Rūja basin. We suppose that the model METQ2007BDOPT calibration results are not the best ones due to lower quality of precipitation data.

Another explanation of the above mentioned calibration differences could be a broad palufied flood plain and a high percentage of wetlands in the rivers Seda and Rūja drainage basins. This reason determines a specific hydrological regime which differs from other studied rivers (i.g., the River Briede) and it is difficult to simulate the rainfall-runoff processes without additional riverbed measurements. As mentioned before, runoff routing of the Lake Burtnieks was simulated on the bases of common hydraulic methods of the reservoirs and open channels. If there was available the channel measurements at the outlet of the lake we could obtained better correlation between simulated and observed water level. Singh V.P. (1995) pointed that all rainfall-runoff models can only be very approximate descriptions of the rainfall-runoff processes and, as such, must be considered to be uncertain in their predictions. However, for many practical purposes, we do not need to include all details in developing a predictive model. Indeed, many successful rainfall-runoff models are essentially very simple. In such models most parameters are not measurable but have to be estimated by calibration using at least some observed runoff data (Seibert 1999).

In this study additionally, the conceptual METQ2007BDOPT model was calibrated and validated to fife different size river basins in Latvia. The results of the model calibration for these river basins showed a good coincidence between the observed and simulated daily discharges from 1961 to 1990: the Nach-Sutcliffe efficiency R^2 varies from 0.86 to 0.50 and correlation coefficient r – from 0.91 to 0.71 (Table 3). The best coincidence was obtained for the Brook Vienziemīte: R^2 – 0.86 and r – 0.91 (Fig. 1). On one hand, it could be explained by the fact that catchment is small (only 5.92 km^2) and used meteorological data fit very well to this drainage area to describe the simulation of hydrological processes. On other hand, we obtained rather good calibration results also for the large river basin such as the River Salaca at Lagaste: R^2 – 0.76 and r – 0.88. The lowest statistical criterions were found for the rivers Imula and the Iecava. The validation of model was done for the next 10-years period from 2000 to 2005, except the river gauging stations Imula-Pilskalni and Iecava-Dupši which was closed since 1995. We obtained lower statistical criterions to compare with calibration period for all gauging stations in this study (Table 3). On one hand, we took different long time periods for the model calibration and validation. In some cases it could be related with data quality of monitoring in 1990's. However, at present it is difficult to explain because the obtained results are preliminary and we need to continue this study.

Table 3. The obtained statistical criterions for fife river basins.

Gauging station	Calibration period (1961-1990)		Validation period (1991-2000)	
	R^2	r	R^2	r
Salaca - Lagaste	0.76	0.88	0.77	0.87
Vienziemīte - Vienziemīte	0.86	0.91	0.63	0.84
Bērze - Baloži	0.72	0.85	0.39	0.80
Iecava - Dupši [1)]	0.58	0.73	0.23	0.79
Imula – Pilskalni[1)]	0.50	0.71	0.43	0.70

[1)] – close since 1995

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MODELLING OF CURRENT AND FUTURE WATER RESOURCES IN NORTHERN EUROPE

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Climate change scenarios generally project increases of the precipitation rates and temperatures in Northern Europe. Changes in the hydrological cycle induced by global warming may affect society profoundly, e.g. with regard to flood and drought risks, changing water availability and water quality. The seasonality may also change, causing new, and sometimes unexpected, vulnerabilities. Studying the effects of climate changes on the hydrological cycle is in itself valuable, and in Northern Europe the amount and timing of available water resources are also of interest when it comes to energy production. In addition, the hydropower sector is profoundly interested in snow amounts, and hydropower producers and managers inquire information about the future snow situation. Given the deregulated power market, hydropower producers are interested in snow amounts at the intra-national level, and for comparison purposes consistent hydrologic modeling over large areas might be desirable.

In order to address some of the above mentioned issues, a large-scale hydrologic model, the Variable Infiltration Capacity (VIC) model (Liang et al., 1994), is implemented for most of Scandinavia, including the Baltic Sea drainage area. The model is run at a scale consistent with global scale hydrological simulations (0.5 degrees latitude by longitude) for current climate and future climate scenarios. Current situation (1961-1990) is compared to simulation results using climate output for two emission scenarios (A1B and B1) for 2071-2100 from the ECHAM5 and HADLEY general circulation models.

The results are used to assess climate impacts on water resources in Northern Europe. The main conclusions, which are consistent with other studies (e.g. Beldring et al., 2006), are that in Scandinavia, the future amount of snow and the length of the snow season decreases, winter runoff increases, and summer runoff decreases. There are, however, fairly large differences between the signal resulting from the two GCM climate outputs and the two emission scenarios. Also, in the southern part of the study area simulated future runoff is lower than simulated current runoff all year round.

In the Nordic countries, water management structures have become increasingly essential to provide water supply, flood control, and electric power. Dams, built for water storage, directly change the dynamics of the water cycle. Exploratory analyses of the effects of man-made reservoirs, which can affect hydrographs substantially, are performed. Seasonally reservoirs can alter downstream streamflow much more than do climate changes, although mean annual flow is only slightly affected.

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RESEARCH AND IMPLEMENTATION OF ALTERNATIVE STORMWATER MANAGEMENT IN THE CITY OF BERGEN, NORWAY 1981 – 2008

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ABSTRACT

This paper describes the implementation of alternative stormwater management (ASWM) in the City of Bergen. When starting the planning of the Bergen South Commercial Parks (BSCP), traditional urban drainage concepts were suggested, increasing the peak flow, Q_{100}, out of the basin from 4.0 to 7.0m^3/s. The landowner denied selling land if the stormwater runoff would increase after urbanization, so new approach to stormwater management had to be introduced. Another challenge was gaps in the knowledge on urban hydrology (UH) and in competence on stormwater management (SWM). The third challenge was that urban hydrological data hardly existed. The project has so far led to the following results:

1. Development of new planning procedures on ASWM, named, "The blue - green concept"
2. Development of new technical solutions, named "The Sandsli – system"
3. Collection of urban hydrological data and development of IDF-curves
4. Building competence in urban hydrology (UH) and ASWM in Norway, by running educational program for Master students
5. Technology transfer of ASWM to other parts of Norway
6. New guidelines for stormwater management in the City of Bergen.

Keywords: alternative stormwater management (ASWM), blue-green concept, infiltration, percolation, stormwater management, Sandsli-system, urbanization

INTRODUCTION

Background

Due to the rapid growth in the new oil related industry in the City of Bergen, the planning process of the Bergen South Commercial Parks (BSCP) started about 30 years ago. BSCP occupies two main watersheds, the Birkeland and Ådland, (fig. 1), totally about 1000 hectares, where commercial, industrial and residential areas are to be built. Traditional urban drainage concepts were suggested, i.e. quick stormwater runoff through pipes and tunnels to downstream areas, increasing the hundred years flow Q_{100} out of the basin from 4.0 to 7.0m^3/s, (Thorolfsson and Skretteberg, 1982) and (Thorolfsson and Sekse, 1999).

Before urbanization the runoff conditions downstream were bad, resulting in flood on agricultural land during the rainy autumn, and low flows in creeks in the dry spring, making irrigation of farm land difficult.

The farmers downstream demanded that the urbanization of BSCP should not make the runoff situation worse than before the urbanization. Instead they demand the developer to use the opportunity to improve the runoff situation by use of new approach for stormwater management, such as detained runoff, local stormwater disposal etc.

The landowners said they would not sell land, if their requirements were not met by the developer, Bergen Housing and Developing company, LtD. Therefore alternative stormwater management (ASWM) concepts had to be implemented to meet the farmer's requirements.

Location, topography, geology and climate

Bergen is located on the west coast of Norway, (Fig. 1). It is the second largest city in Norway with a population of 245.000 inhabitants in January 2007. The research area covers BSCP that includes the watersheds Birkeland with the build-up areas, Kokstad and Sandsli and Ådland watershed with Lønningen. Large office buildings for

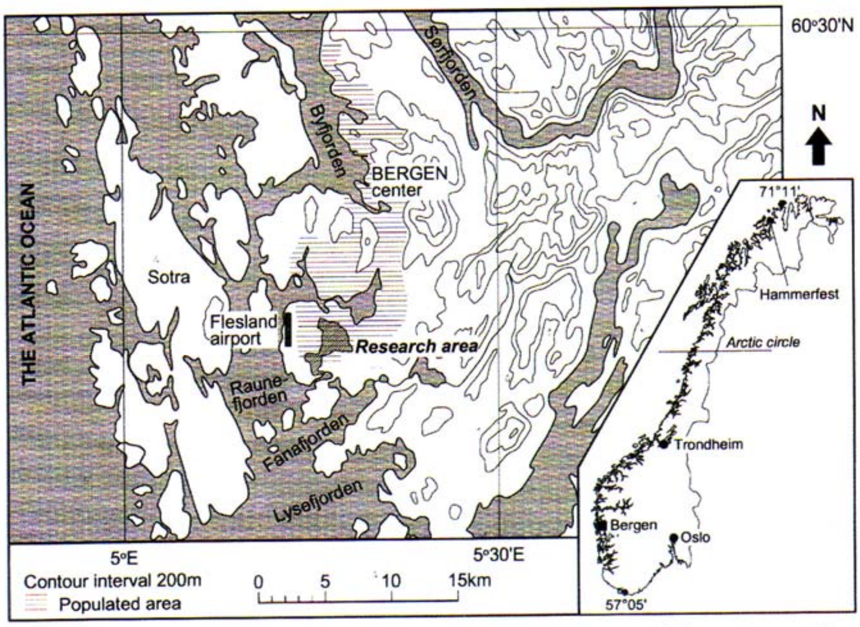

Figure 1. Location of Bergen, bottom right, and the research area in the watersheds, Birkeland and Ådland

companies such as Statoil, Norsk Hydro and Telenor etc. are located here having large impervious areas, such as the Statoil research building has a roof on 2000 square meters.

The main feature of the landscape in the Birkeland watershed is three lakes; Skrane Lake, at +40.50 masl, Birkeland Lake, at + 36.00 masl and Håvardstun Lake, at +28.00, interconnected by creeks (fig 2.). The ground conditions may be characterized by rocky ground with a shallow layer of sandy soil or a thicker layer of boggy ground over the bedrock. The rock has often week zones with crushed stones. During development thick layers of boggy ground may be removed and replaced with crushed rock and gravel.

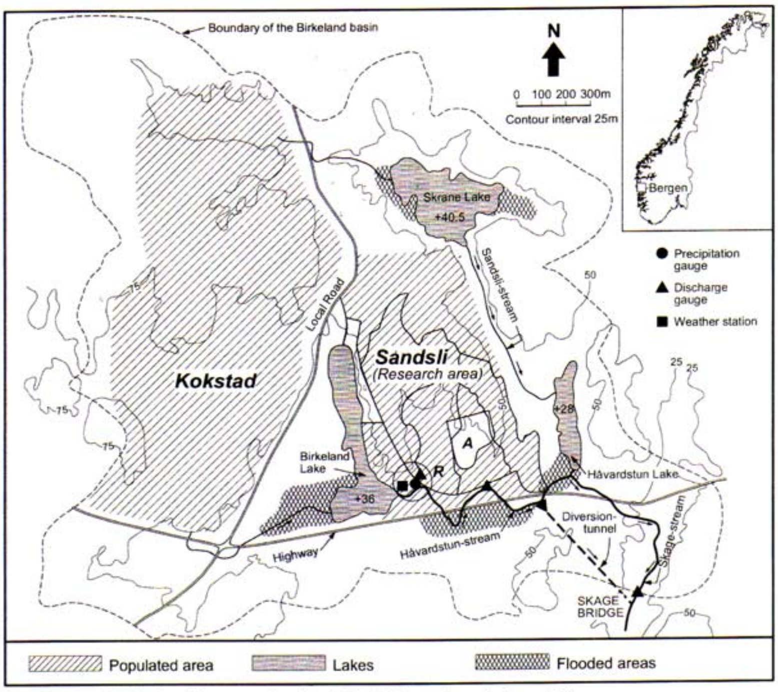

Figure 2. Birkeland watershed with Kokstad and Sandsli

The climate in Bergen is a maritime temperate climate (Cfb) according to Køppen’s classification, Køppen (1918/1936). Winters are relatively warm, while the summers are cool. Bergen is one of the warmest cities in Norway, thanks to the Gulf Stream. 10°C and rain can happen both in January and July.

In Bergen center the annual average precipitation is 2250 mm/yr, among the highest in Europe, with highest precipitation in autumn. This is resulting in huge urban runoff volumes that have to be managed. At Bergen airport, Flesland only 1 – 4 km west of BSCP, the precipitation is 1850mm/yr.

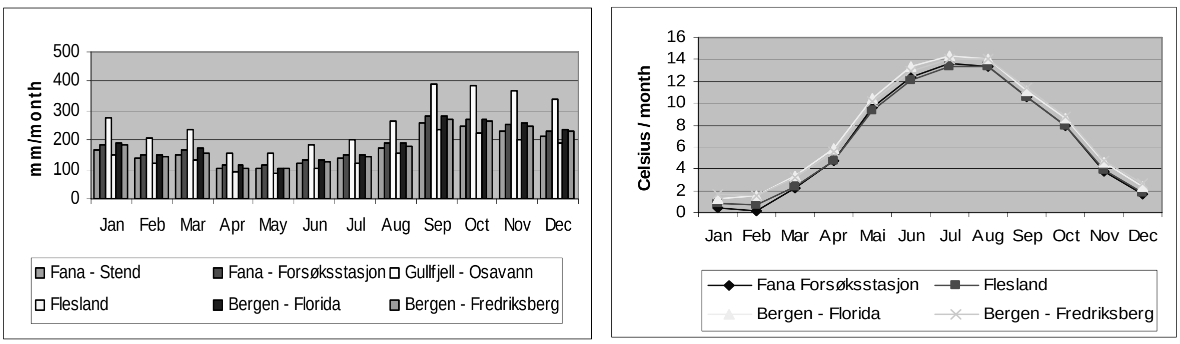

Figure 3. Average monthly precipitation and air temperatures in some locations in Bergen.

In the beginning planners and engineers were facing big gaps in the knowledge on urban hydrology (UH) and in competence on stormwater management (SWM). Another problem was that relevant urban hydrological data hardly existed. Therefore a long-term project to close these gaps was initiated, and a project group was established to coordinate the activities. This group was named PUB (Project group on Urban hydrology in Bergen

NEW APPROACH TO STORMWATER MANAGEMENT

When seeking for new approach to stormwater management, planners and engineers analyzed the topography and the landscape and found it suitable to use the lakes and the creeks as an open stormwater management solution. It could give stormwater detention and delayed runoff out of the watershed, but the lakes and banks along creeks had to be regulated to areas for stormwater management.

This solution was named, the blue - green concept, because of the blue and green surface to be used. Areas needed for the blue - green concept had to be regulated as stormwater management areas. This was the first time to do so in the City of Bergen and even in Norway. Some officials were skeptical to that solution, but the government of the City approved the revised plan including areas for stormwater management. The government also stated that these areas should be kept as untouched as possible, because of biological diversity and recreation.

This was a new situation for bureaucrats, engineers and practitioners. Instead of laying stormwater pipes and building tunnels, the task became how to manage stormwater runoff according to the blue - green concept. Some of they became skeptical and even boycotted this solution, making it difficult implementing it.

Some alternatives were considered, but the following solution was selected; Skrane lake was regulated by 0.50 meters, Birkeland lake was regulated by 0.67 meters and Håvardstun Lake regulated by 0.96 meters and the banks around lakes and creeks were allowed to be flooded, as shown in (fig. 2) and the three creeks, Sandsli stream, Håvardstun stream and Skage stream are to be kept untouched as possible. This solution was considered to keep flow at Skage Bridge around 4.00m^3/s i.e. as before urbanization. Furthermore, a planned stormwater tunnel at a cost of NOK 6 million was found not to be needed, but it was build for security reason, while officials did not thrust the proposal because of the uncertain data and not existing experiences with that kind of stormwater solutions.

The blue-green concept was aimed to reduce the peak flow to downstream agricultural areas, but it did not solve the low flow problems in the spring. Infiltration and percolation of the stormwater into the ground would recharge the groundwater reservoir and result in slow subsurface flow to downstream areas.

Therefore, the second project was on development of measures that would recharge the groundwater. The first idea was to put as much as possible of the stormwater from roofs into the ground and may be also some of the stormwater from light traffic surfaces. It resulted in development and later refinements of the Sandsli-system, Thorolfsson (1999a, 1999b and 1999c). The Sandsli – system was first implemented in the Sandsli catchment in 1981, therefore the name Sandsli-system.

The Sandsli-system is an integrated sanitary and stormwater management system, described by Thorolfsson (1997, 1998, 1999a and 1999b). When excavating for road and house foundations, water and sewer pipe trenches etc., the impermeable top layer is removed and replaced with blasted stones, when broken stones may be filled in deep holes where boggy ground has been removed down to the bedrock. Under house foundations and in pipe trenches crushed stone and gravel are filled, giving 30% porosity. These fillings may be used for detention and percolation into cracks and weakness zones in the bedrock.

The effectiveness of the Sandsli-system has been documented in some research projects, Thorolfsson (1999 and 1997), and it was especially documented under the heavy storms on 14th September and 14th November 2005. The system has been used in other locations in Norway and in Vestmannaeyjar in Iceland, Thorolfsson (1999) and Aasen et. al (2005).

The third goal of the project was to solve the lack of reliable urbanhydrological data. Therefore Sandsli urban hydrological measuring station was established in august 1982, recording precipitation, stormwater runoff, and air temperature and snowmelt rate with high time resolution, i.e. minutes. After 26 years in operation IDF-curves have been developed, Thorolfsson et. al (2007).

The fourth goal of the project was on lift the competence within stormwater management in the Bergen region, by supporting education and training program for Master students in the Civil and Environmental Engineering at NTNU, Norway. Eighteen students have passed the Master degree based on the above mentioned projects; among they are the present Director of the Water supply and Wastewater Department of the Municipality of Bergen, Aasen and Sekse (2005).

Based on the results of the above mentioned projects, new guidelines for Stormwater Management in the Municipality of Bergen, has been developed, Bergen City (2005).

LESSIONS LEARNED

This project has realized that planning the use of new concepts for stormwater management it is important to inform properly on the system to be used and even educate the people involved, especially bureaucrats, engineers and practitioners involved. This is to avoid unnecessary skepticisms that may result in improper technical solutions, which are not functioning well. But if the solutions are properly implemented they are functioning well. It seem that both planning, projecting and especially implementation of alternative stormwater management solutions need more skilled people and more effort has to be put into the whole process, that using traditional stormwater solutions.

The blue-green concept and the Sandsli systems were implemented in BSCP due to the need for flood protection downstream and to increase groundwater recharge. They were not aimed for pollution control so location of unclean industry in these areas must be restricted.

Stormwater will always contain some impurities, even stormwater from roofs are affected by bird skit etc. These impurities will find the way into the ground or into the nearest watercourse, if the stormwater is conveyed to such recipients and be may accumulated. Investigations shows, that the lakes in BSCP have been affected by such discharge. Today, the water quality in Birkeland Lake is poor, i.e. water class D, while the water quality in Skrane Lake and Birkeland Lake is, not good, i.e. water class C. This is reducing the recreational value of the lakes. Some clean up measures have to be taken.

The lesion learned is when planning to use alternative stormwater managements, the stormwater pollution has to be considered and removed before it enters the system, to avoid degradation of nearby watercourses or the ground.

The maintenance of the systems must by carefully follow up and clear guidelines prepared. It seems to be a widely spread misunderstandings that these systems need less maintenance the traditional systems. It is opposite. The experiences are that they need to be more maintained.

SUMMARY AND RESULTS

When starting the urbanization of huge area of Bergen South Commercial Parks (BSCP), a serious lack in knowledge on urban hydrology (UH) and stormwater management (SWM) in the Bergen region became apparent. There were lack of planning and projecting methods and also appropriate models and urban hydrological data were missing

This situation initiated a long-term research project, where four goals were identified. All together they were directed to create a basis for better, cheaper and more correct stormwater management and dimensioning of stormwater systems in the Bergen region in Norway. The results may be summarized as follows:

1. Development of new planning procedures on ASWM i.e. “The blue – green concept” and “Planning flood ways” in urban areas

2. Development of new technical solutions on ASWM, i.e. “the Sandsli – system” and “stormwater solutions without pipes”
3. Collection of urban hydrological data and developing of IDF-curves
4. Building competence in UH and ASWM in Norway, i.e. educational program for Master students
5. Technology transfer in UH and in ASWM to other parts of Norway.
6. New guidelines for Stormwater Management in the City of Bergen

The project has been led by the Project group for Urban hydrological activities in Bergen (PUB), established in April 1981. PUB has been active since 1981 and has often been the only environment in Norway doing investigations and research on SWM. PUB has promoted education of seventeen Civil Engineers. Some of them have become leading persons within stormwater management in Norway and particularly in the City of Bergen. The leader for PUB from the start to data has been Associate Professor Sveinn T. Thorolfsson. He is the driving force in PUB, still seeking for new challenges within alternative stormwater management (ASWM).

ACKNOWLEDGEMENTS

The authors thank the Norwegian University of Science and Technology (NTNU), the City of Bergen, the Bergen Housing and Development Company, and other supporters. Gratitude goes also to the Norwegian Research Council (NFR), Norwegian Hydrological Committee (NHK) and Norwegian State Pollution authority (SFT) for their supports. We send also our gratitude to all the Master students, who have contributed. Finally, thanks to the former chief for the Water Supply and Wastewater Department in Municipality of Bergen, Alf Rustad, for his initiative to start the project in 1981.

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SESSION 4: UNCERTAINTY AND EXTREMES IN HYDROLOGY

DOWN-SCALING ATMOSPHERIC PATTERNS TO MULTI-SITE PRECIPITATION AMOUNTS IN SOUTHERN SCANDINAVIA

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ABSTRACT

Non-homogeneous Hidden Markov Models (NHMMs) are applied for down-scaling atmospheric synoptic patterns to winter multi-site daily precipitation amounts. The implemented NHMMs assume precipitation to be conditional on a hidden weather state that follows a Markov chain, whose transition probabilities depend on the atmospheric information. The gridded atmospheric fields are summarized through the Singular Value Decomposition (SVD) technique. SVD is applied to geopotential height and relative humidity at several pressure levels, to identify their principal spatial co-varying patterns with precipitation. We assume the common hidden weather state process to completely account for the spatio-temporal structure of precipitation: given the current weather state, precipitation occurrences are assumed to be conditionally independent Bernoulli random variables, both temporally and spatially; and precipitation amounts are modelled as spatially independent Gamma distributed random variables. This modelling approach is applied to 51 precipitation gauges in Denmark and southern Sweden, over the period 1981-2004. The downscaling model produces robust predictions of data statistics such as expected precipitation amounts and spell duration distributions. Moreover, the model-defined weather states show a satisfactory degree of

physical consistency. Finally, possible advancement is outlined, especially regarding the reproduction of space-dependency.

INTRODUCTION

The interest in modelling precipitation arises from the need for input series to be used in hydrological models. Recent predictions of global climate change have caused an increasing interest in assessing effects at regional scale and precipitation is a key factor in the description of hydrologic regimes. Historically the modelling efforts have followed two main lines: some models include physical laws, while others concentrate on the statistical description of the records. The formulation of NHMMs originates from the frameworks of stochastic precipitation modelling and of downscaling, to simulate time series that are consistent with atmospheric historical records and with outputs of General Circulation Models (GCM) (Hughes et al., 1994).

GCMs have been used for predicting possible changes in the climate due to increasing CO_2 emissions (Hughes et al., 1993). As GCMs typically simulate climatic variables on 3^o latitude $\times$ 3^o longitude grids or similar, they are not suitable to predict non-smooth fields such as precipitation (Hughes and Guttorp, 1994). Previous studies, including Giorgi and Mearns (1991), and Bates et al. (1998), have pointed to the need for models that can downscale the simulations of GCMs, as well as historical synoptic-scale atmospheric records, to local-scale precipitation patterns.

Giorgi and Mearns (1991) nested a finer mesh Limited Area Meteorological model (LAM) in a GCM, which provided the boundary conditions for the higher resolution modelling. This methodology involved high computational costs and the biases produced by the GCM were not attenuated by the LAM.

Although traditional stochastic approaches did not include atmospheric information as a conditioning factor, the attractiveness of Markov chains for modelling precipitation or climate processes has long been recognized, see for example Green (1970), Haan et al. (1976), and Zucchini and Guttorp (1991).

Weather State Models (WSMs) constitute another approach to downscaling. The concept of WSM was developed by Hay et al. (1991), and is based on assigning the observed days to weather types that can be defined through expert meteorological knowledge (Bardossy and Plate, 1991) or automatic classification methods (Hughes et al., 1993). After these assignments, precipitation distributions are estimated for each weather class.

NHMMs for precipitation were developed by Hughes and Guttorp (1994), Hughes et al. (1999), and Bellone et al. (2000). As for WSMs, NHMMs can simulate precipitation conditioned on atmospheric information. However, in NHMMs the weather types are described as states of a Markov chain and they are not defined a priori by classifying atmospheric patterns. Weather classes are instead identified as precipitation probability patterns, while the atmospheric variables influence the state transitions.

NON-HOMOGENEOUS HIDDEN MARKOV MODELS

Model formulation

NHMMs assume the observed precipitation process to be driven by an unobserved weather state process. The weather state at time *t* is represented by the discrete stochastic variable s_t following a first order Markov chain, where the transition probabilities depend on the current atmospheric information. At each time step *t*, the probability distribution of the precipitation pattern is assumed to be uniquely determined by the weather state prevailing at *t*. Defining $\underline{r}_t$ and $\underline{x}_t$ as the precipitation pattern and the atmospheric information at time *t*, respectively, the assumptions outlined above can be written as:

$$pmf(s_t) = f(s_{t-1}, \underline{x}_t) \tag{1}$$

$$pdf(\underline{r}_t) = g(s_t) \tag{2}$$

where *pmf* means probability mass function. Given assumption (1), the chosen parameterisation for the unobserved process is (Hughes and Guttorp, 1994):

$$\Pr\left[s_t = j \middle| s_{t-1} = i, \underline{x}_t\right] \propto p_{ij} \exp\left[-\frac{1}{2}\left(\underline{x}_t - \underline{\mu}_{ij}\right)' \underline{\underline{V}}^{-1}\left(\underline{x}_t - \underline{\mu}_{ij}\right)\right] \tag{3}$$

where $\underline{\underline{V}}$ is the covariance matrix of the atmospheric variables, p_{ij} is a base-line transition probability from state *i* to *j*, and $\underline{\mu}_{ij}$ can be interpreted as the mean values of the atmospheric variables when the current state is *j* given that it was *i* at the previous time step.

The observed process, following assumption (2), is parameterised according to Bellone et al. (2000): given the current state, $pdf(\underline{r}_t)$ is defined as a mixture of a probability mass for the interval [0,*c*] and a Gamma distribution:

$$\Pr\left[\underline{r}_t = \underline{r} \middle| s_t = j\right] = \prod_{g=1}^{G} \left[\o_{gj} Ga\left(r_g - c;\quad \kappa_{gj}, \varphi_{gj}\right)\right]^{\delta(r_g - c)} \left(1 - \o_{gj}\right)^{1-\delta(r_g - c)} \tag{4}$$

where $\delta(\bullet)$ is the Dirac delta function, *G* is the number of rain gauges, $\o_{gj}$ is the precipitation occurrence probability at gauge *g* for state *j*, $Ga(\bullet;\quad \kappa_{gj}, \varphi_{gj})$ is the Gamma *pdf* with state- and gauge-specific parameters κ_{gj} and θ_{gj}, and *c* is the threshold below which precipitation is neglected, fixed to 0.2 mm. A further assumption is underlying (4): precipitation occurrence probabilities at different gauges are conditionally independent. Thus the common weather state, modelled by the Markov chain, entirely accounts for the spatial and temporal dependencies of the observed process.

Atmospheric variables

Synoptic-scale atmospheric fields are available on regular grids. As the study area spans several nodes, a summarizing method is used to reduce the gridded fields into a few explanatory variables and to limit the model dimension. Here we follow the approach of Bellone et al. (2000) that is based on Singular Value Decomposition (SVD). For each standardised atmospheric field $\underline{z}$ we build a matrix $\underline{\underline{A}}$, whose element a_{ig} is the covariance between the standardised atmospheric field at node i and the standardised precipitation process at gauge g. We can decompose $\underline{\underline{A}}$ through SVD and obtain:

$$\underline{\underline{A}} = \underline{\underline{V}}\,\underline{\underline{U}}\,\underline{\underline{H}}' \tag{5}$$

The summary variable x_i is obtained by multiplying $\underline{z}$ by the i^{th} column of $\underline{\underline{V}}$. It explains the fraction $u_i^2 \Big/ \sum_{j=1}^{G} u_j^2$ of the covariance between $\underline{z}$ and the precipitation process, where u_i are the singular values of $\underline{\underline{A}}$.

Parameter estimation

Maximum likelihood estimates (MLEs) of model parameters are derived through the iterative Expectation-Maximization (EM) algorithm (Hughes et al., 1999). Each iteration n of the EM algorithm consists of two steps. The first step (Expectation) computes the expected value of the model log-likelihood function given the previous parameter estimates $\hat{\theta}^{(n-1)}$:

$$\begin{aligned}
&Q\left(\theta \middle| \underline{r}_1^T, \underline{x}_1^T, \hat{\theta}^{(n-1)}\right) = E\left[\log L(\theta) \middle| \underline{r}_1^T, \underline{x}_1^T, \hat{\theta}^{(n-1)}\right] = \\
&\sum_{i=1}^{S}\sum_{j=1}^{S}\sum_{t=2}^{T} \Pr\left[s_{t-1}=i, s_t=j \middle| \underline{r}_1^T, \underline{x}_1^T, \hat{\theta}^{(n-1)}\right] \log\left\{p_{ij}^{(n-1)} \exp\left[-\frac{1}{2}\left(\underline{x}_t - \underline{\mu}_{ij}^{(n-1)}\right)' \underline{\underline{V}}^{-1}\left(\underline{x}_t - \underline{\mu}_{ij}^{(n-1)}\right)\right]\right\} + \\
&\sum_{j=1}^{S}\sum_{t=2}^{T} \Pr\left[s_t=j \middle| \underline{r}_1^T, \underline{x}_1^T, \hat{\theta}^{(n-1)}\right] \log\left\{\prod_{g=1}^{G}\left[\o_{gj}^{(n-1)} Ga\left(r_{gt}-c;\ \kappa_{gj}^{(n-1)}, \varphi_{gj}^{(n-1)}\right)\right]^{\delta(r_{gt}-c)}\left(1-\o_{gj}^{(n-1)}\right)^{1-\delta(r_{gt}-c)}\right\}
\end{aligned} \tag{6}$$

where S is the number of weather states; T is the record length; $\underline{r}_1^T$ and $\underline{x}_1^T$ are, respectively, the measured rainfall patterns and the atmospheric variables from time 1 to T; and the state probability terms can be computed by the Forward-Backward algorithm (Rabiner and Juang, 1986; Akintug and Rasmussen, 2005).

The second step (Maximisation) maximises $Q\left(\theta \middle| \hat{\theta}^{(n-1)}\right)$ with respect to θ, obtaining the new estimates:

$$\hat{\theta}^{(n)} = \arg\max_{\theta} Q\left(\theta \middle| \hat{\theta}^{(n-1)}\right) \tag{7}$$

As shown in (6), the two terms summing to Q involve different parameters. Thus they can be maximised individually using the Lagrange multipliers method, enforcing constraints $\sum_{j=1}^{S} p_{ij} = 1$ and $\sum_{j=1}^{S} \underline{\mu}_{ij} = \underline{0}$ (for $i = 1,\ldots,S$).

The EM steps are iterated until a maximum of the likelihood function is reached. However, the likelihood of a NHMM depends on a large number of parameters and may have several local maxima. Therefore the choice of initial parameter values is essential for obtaining proper estimates (Bellone et al., 2000). The EM algorithm is initialised by applying the convergent variant of MacQueen's *k*-means clustering method (Anderberg, 1973) to the *G*-dimensional time series $\{\underline{r}_t\}$.

Model selection

The choice of model dimension involves deciding how many weather states are to be defined and which atmospheric covariates are to be included. This means trading off between goodness-of-fit and parameter parsimony. We use the Bayesian information criterion (BIC) (Kass and Raftery, 1995) as a decision factor. The BIC appears to yield reasonable models although some assumptions underlying its definition are violated by applying it to NHMMs (Hughes et al. 1999). Defining D as the number of free parameters, the BIC is given by:

$$BIC = -2\log L\left(\hat{\theta}_{MLE} \middle| \underline{r}_1^T, \underline{x}_1^T\right) + D\log(G \cdot T) \tag{8}$$

DATA

Daily precipitation and atmospheric data were extracted for the period 1981-2004 for the months of November through February. The decision to consider only autumn-winter periods is due to the recognised difficulties of associating spring and summer precipitation with synoptic-scale atmospheric patterns (Linderson, 2000). NHMMs are indeed not likely to give reasonable results for seasons and regions where convection drives precipitation: convective processes are characterized by small spatial scales and may not be predictable by synoptic atmospheric information (Hughes et al., 1999).

Precipitation measurements were made available by DMI (Danish Meteorological Institute) and SMHI (Swedish Meteorological and Hydrological Institute). Data from 51 gauges, located in Denmark and southern Sweden, were used and their positions are shown in Fig. 1. Missing observations have a relatively low frequency (0.75%), hence we decided to proceed with data-filling by multiple linear regression.

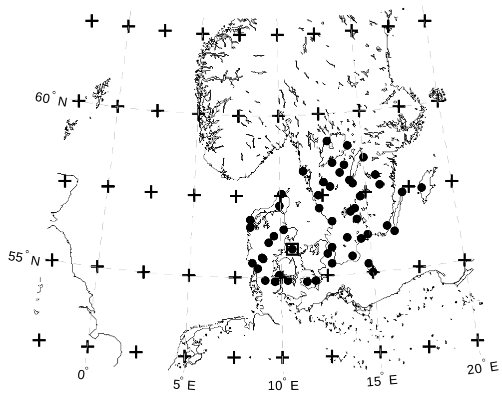


Figure 1. Location of precipitation gauges (•) and atmospheric grid nodes (+).

The atmospheric fields are NCEP Reanalysis Data provided by the NOAA-CIRES Climate Diagnostics Center (Boulder, Colorado, U.S.A.). Sea level pressure, geopotential height, and relative humidity, at several pressure levels, have been chosen because of their relevance to the precipitation phenomenon: geopotential height and sea level pressure represent the circulation patterns, while relative humidity indicates the degree of saturation of the atmospheric layers. No missing observations were reported for these data. The atmospheric grid nodes (spaced 2.5^{o} in latitude and longitude) used for this application are plotted in Fig. 1.

RESULTS

The records were split into calibration (1981-1996) and validation (1997-2004) data sets. We chose the best model by comparing setups including 2 or 3 atmospheric covariates and defining from 5 to 11 weather states. BIC scores indicated a group of best performing models, but its value had relatively small variations among these setups. Hence, the final choice was made according to other considerations. We wanted to include at least one covariate accounting for synoptic circulation and one for air humidity, but we also wanted to avoid redundancies in the atmospheric input and in definitions of states. Thus we chose a NHMM with 8 weather classes, downscaling geopotential height at 1000 hPa and relative humidity at 850 hPa.

Precipitation statistics

Here we present model-based reproductions of some precipitation statistics, for both calibration and validation periods. We chose the gauge at Tranebjerg (marked with a square in Fig. 1) to show the gauge-specific results, namely the spell duration distributions and the precipitation quantiles. As overall statistics, mean precipitation amounts and lag-0 cross correlations are reported in Fig. 2 and Fig. 3.

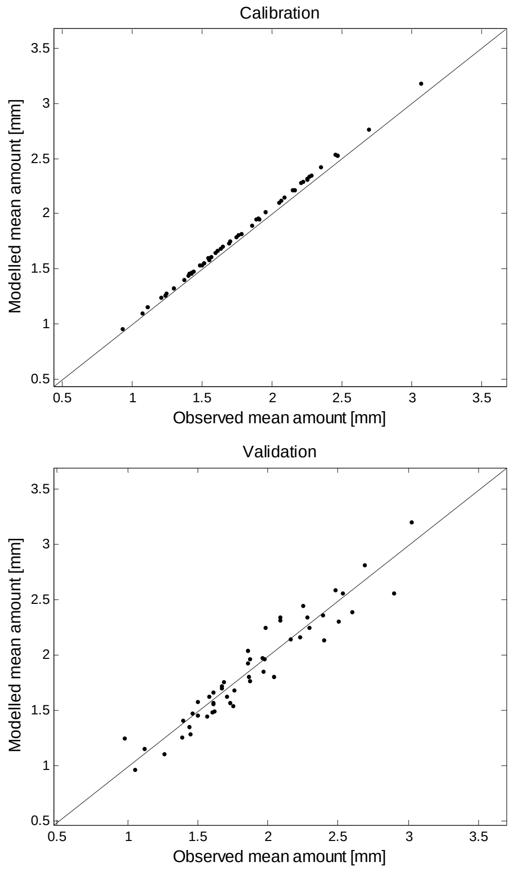


Figure 2. Precipitation mean amounts.

The observed mean precipitation amounts are reproduced with good precision in the calibration although a small bias can be seen. As expected, model-based values for calibration have slightly lower precision, but the bias remains small. Lag-0 cross correlations are measures of the linear dependence between precipitation at each pair of gauges at the same time. Their reproduction is relatively unbiased for values up to 0.4 approximately, while higher cross-correlations are systematically underestimated by the model. This tendency is observable for both calibration and validation results and highlights the limitation of assuming conditional spatial independence of the precipitation process. This limitation was also noted by Bellone et al. (2000).

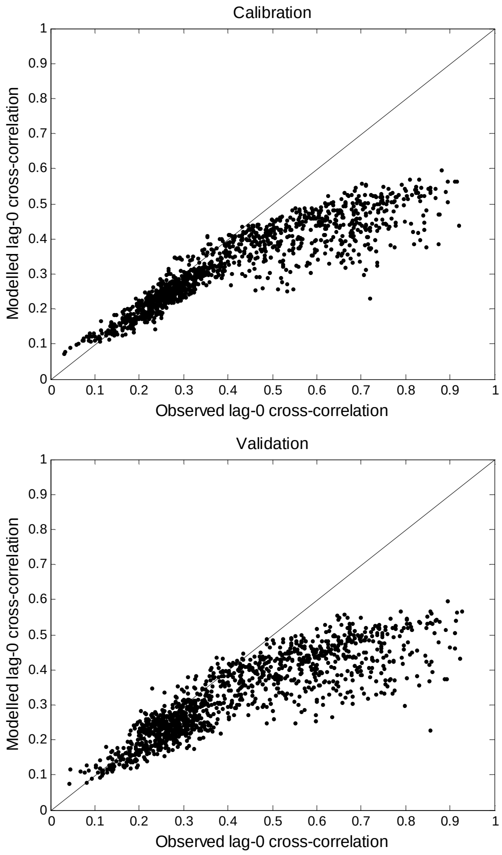


Figure 3. Lag-0-correlations.

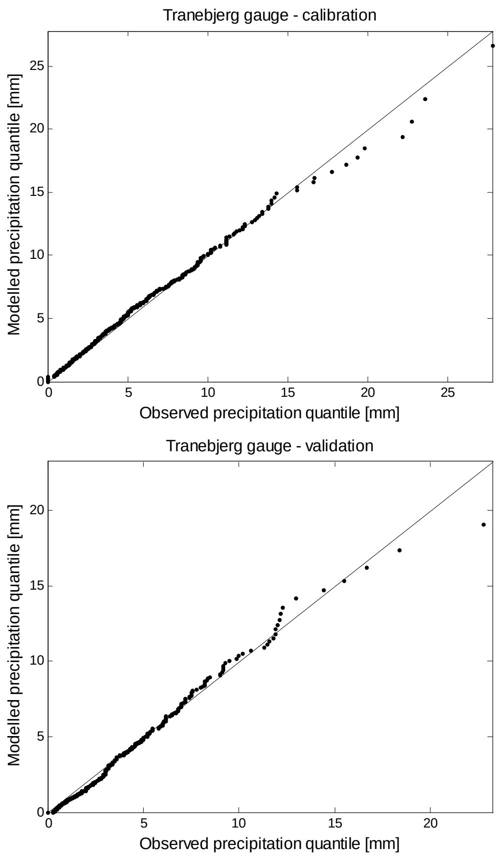


Figure 4. Precipitation quantiles (at Tranebjerg).

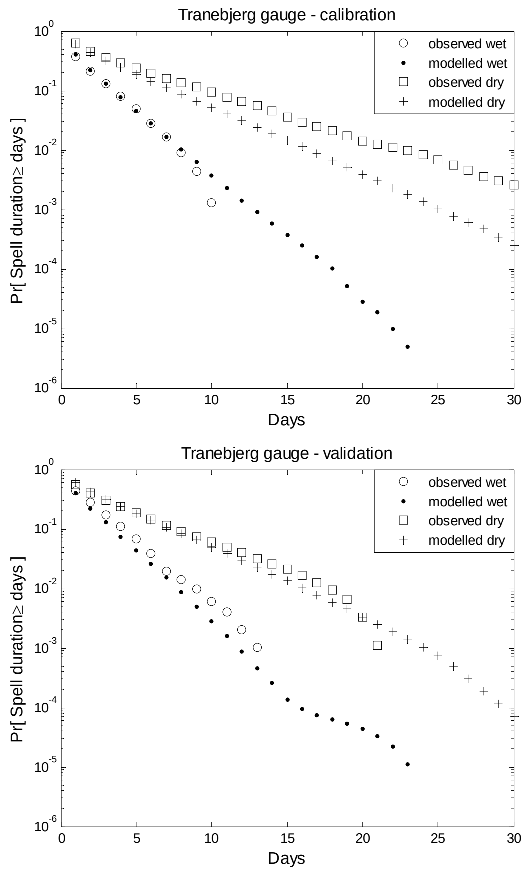
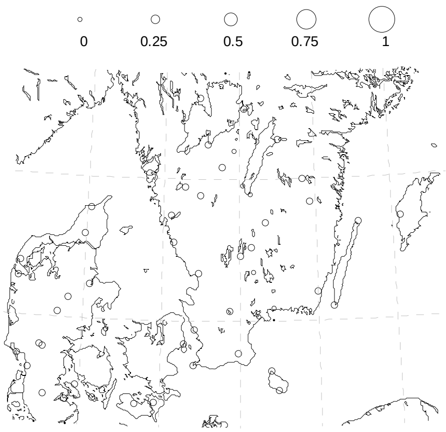


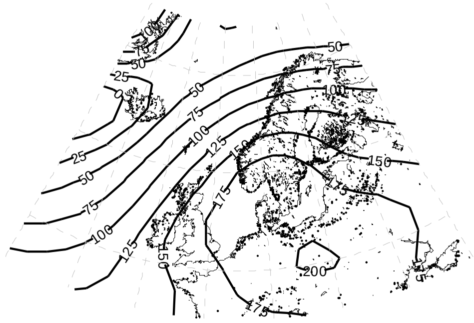
Figure 5. Dry- and wet-spell duration distributions (at Tranebjerg).

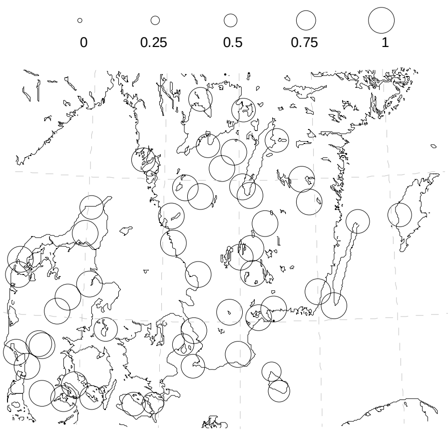
Fig. 4 shows observed versus modelled values of the precipitation quantiles for the gauge at Tranebjerg. The good reproduction during both calibration and validation suggests that the Gamma distribution is well suited for this application. The reproduction of both dry- and wet-spell distributions is satisfactory under calibration as well as validation, although there is a slightly downward bias. However, the bias is small, suggesting that the assumption of temporal conditional independence is acceptable for reproducing the generally low persistence of the precipitation process in this region. Indeed, auto-correlation coefficients (for any lag) were found to be lower than 0.2 for most gauges.

Weather states

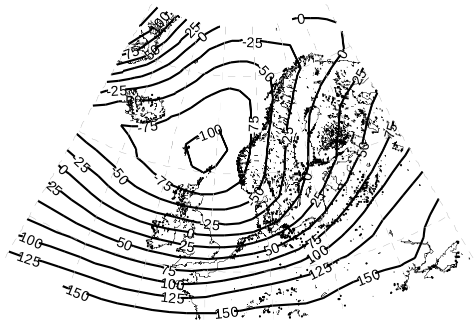


Geopotential height at 1000 hPa - averaged values [m]





Geopotential height at 1000 hPa - averaged values [m]



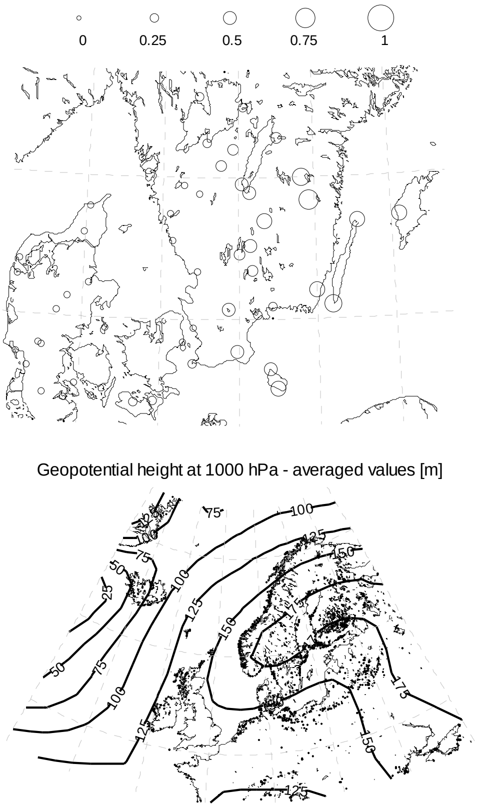


Figure 6. Precipitation occurrence probabilities (left) and averaged geopotential height at 1000 hPa (right). From top to down: states 1, 2 and 3.

The model-defined weather classes allow further validation of the model by checking whether the precipitation patterns are physically consistent with their corresponding averaged synoptic atmospheric patterns. Here we present a comparison between precipitation occurrence probabilities and geopotential height at 1000 hPa (GH1000) for three weather states (Fig. 6).

Weather state 1 is characterized by very low precipitation probability at all gauges. The GH1000 pattern depicts blocking conditions, with a strong high pressure system over continental Europe and southern Scandinavia. Hence, state 1 corresponds to the typical high-pressure weather. State 2 corresponds to a westerly weather driven by a deep low pressure approaching Scandinavia. Precipitation probabilities are high at all the gauges. Averaged GH1000 values suggest that winds blow from south-west, carrying precipitation over the whole region. For state 3, rainfall probabilities are relatively high over the easternmost part of the region and progressively decrease westwards. The SLP pattern shows a high pressure system over Scandinavia and Russia, corresponding to an easterly regime. Under these conditions, air masses move from east and get heated and moisturized by the Baltic Sea. Thus precipitation is most likely to occur over the eastern part of the region.

DISCUSSION AND CONCLUSIONS

NHMMs adequately predicted key statistics such as expected precipitation amounts and spell length durations for both the calibration and the validation periods. Spatial correlations were predicted with somewhat lower accuracies and systematic biases caused by certain model limitations.

The validation results highlight the potential of NHMMs for producing realistic simulations under slightly altered climate scenarios. However, the ability of NHMMs to downscale climate change scenarios cannot be proved on the basis of the presented results. The good physical significance of the weather classes is encouraging in this aspect. Although no direct weather classifier was used for identifying the states, the associated averaged atmospheric patterns showed reasonable correlations with the precipitation probabilities defining the weather states. Moreover, these correspondences provided insight into the atmospheric processes that drive precipitation over the region.

The physical consistency of the weather classes suggests that the SVD-based summarising technique succeeds in capturing the principal co-varying spatial patterns of the atmospheric fields. This result is supported by the good predictions during validation.

The limited reproduction of cross-correlations highlights the lack of an explicit parameterization of the spatial dependence structure of precipitation. As concluded by Hughes et al. (1999), the hypothesis of conditional spatial independence proves not to be suitable for spatially dense networks of gauges such as the one studied here. Future model developments should focus on a spatial model for the precipitation amount distributions.

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DYNAMICAL DOWNSCALING OF PRECIPITATION – PART I: COMPARISON WITH GLACIOLOGICAL DATA

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ABSTRACT

Atmospheric flow over Iceland has been simulated for the period January 1961 to July 2006, using the mesoscale MM5 model driven by initial and boundary data from the ECMWF. The simulated precipitation is compared to estimates derived from mass balance measurements on the Icelandic ice caps. It is found that the simulated precipitation compares favourably with the observed winter balance, in particular for Hofsjökull, where corrections to take liquid precipitation and / or winter ablation into account have been made, and for the comparatively high altitude outlet glaciers Dyngjujökull and Brúarjökull, where such corrections are relatively unimportant.

INTRODUCTION

The geographical distribution of precipitation in Iceland is poorly known but very important for hydrological applications, both in general and particularly in the context of climate change. Therefore, an extensive task carried out in the recent VO project (Jóhannesson et al., 2007) was concerned with modelling of precipitation and a compilation of precipitation data sets on a regular grid covering the whole country. These data sets provide the opportunity to model river runoff and glacier mass balance both in the current climate and also in a hypothetical future climate based on the CE / VO climate change scenarios.

The climate of Iceland is largely governed by the interaction of orography and extra-tropical cyclones, both of which can be described quite accurately by present day atmospheric models. As a result, dynamical downscaling of the climate, using physical models, gives valuable information about precipitation distribution, especially in the data-sparse highlands.

In this paper we compare dynamical downscaling of large-scale meteorological fields provided by the ERA40 reanalysis (Uppala et al., 2005) to precipitation estimates derived from mass balance measurements on the Icelandic ice caps. The dynamical downscaling is done by using the mesoscale MM5 model (Grell et al., 1995).

This paper begins with a description of the model approach, followed by comparison of the model results to data and concluding remarks.

MODELLING WITH THE MM5 MODEL

The idea of using limited area models (LAMs) for regional climate simulations was introduced by Dickinson et al. (1989) and refined by Giorgi (1990). One of the benefits of such an approach is that it is relatively inexpensive in terms of computer resources used for simulations of the atmospheric flow at relatively high spatial and temporal resolutions. As resolution is increased, processes governed by the interaction of the large scale flow and topography become better resolved by the models. One drawback of this approach which is not present in global climate models is that the simulations are dependent on the lateral boundary conditions. These can constrain the model dynamics and hence affect the results (*e.g.* Warner et al., 1997). To minimize the constraining effects of the boundary conditions, Qian et al. (2003) suggested consecutive short term integration, overlapping in time as to minimize the effects of spin-up, instead of a single long term integration. Other investigators (*e.g.* Giorgi and Mearns, 1999) opt for longer integration times, emphasising the importance of the model to be free to develop its own internal circulations. Liang et al. (2004) used this approach when simulating precipitation over the U.S. during 1982–2002 using the MM5-based regional climate model CMM5.

Several case studies investigating orographic forcing of precipitation have been made in recent years. Chiao et al. (2004) used the MM5 model at a 5 km horizontal resolution to simulate a heavy precipitation event during MAP IOP–2B. The precipitation was satisfactorily reproduced by the model although the total amount of precipitation was slightly higher than measured by rain-gauges. Buzzi et al. (1998) simulated a 1994 flooding event in northwestern Italy. The role of orography was found to be crucial in determining the precipitation distribution and amount.

Atmospheric flow over Iceland was simulated for the period January 1961 through June 2006 using version 3–7–3 of the PSU/NCAR MM5 mesoscale model (Grell et al., 1995). The domain used is 123×95 points, centered at

64° N and 19.5° W, with a horizontal resolution of 8 km. There are 23 vertical levels with the model top at 100 hPa and model output is every 6 hours. The domain setup is shown in Figure 1. The MM5 model was used with initial

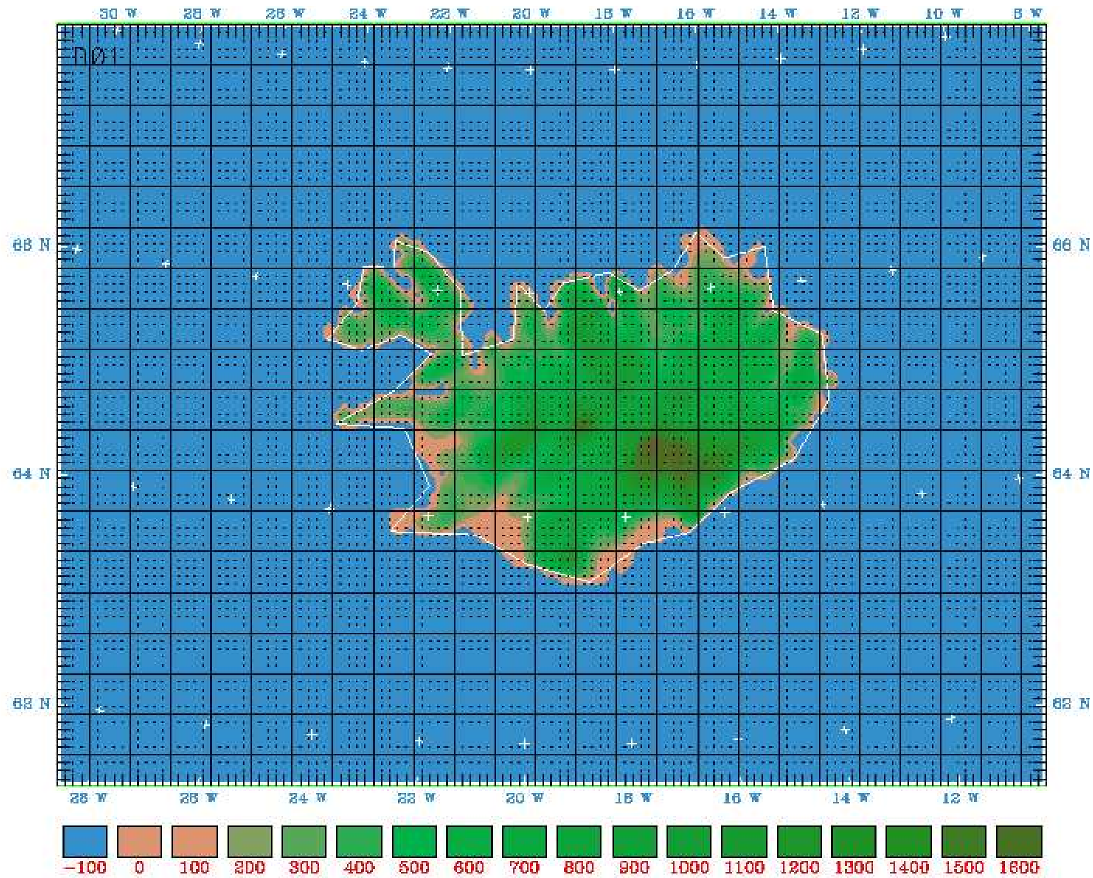

Figure 1: Domain setup of the MM5 model, horizontal grid size is 8 km.

and lateral boundaries from the ERA40 re-analysis project to 1999. After that date, operational analysis, from the ECMWF were used. The ERA40 data were interpolated from a horizontal grid of 1.125° to 0.5° prior to being applied to the MM5 modelling system. The modelling approach differs from that used by Bromwich et al. (2005). Instead of applying many short term (*i.e.* on the order of days) simulations and frequently updating the initial conditions, the model was run over a period of approximately six months with only lateral boundary conditions updated every six hours. This was made possible by taking advantage of the OSU land surface model (Chen and Dudhia, 2001).

VERIFICATION OF SIMULATED PRECIPITATION

Rögnvaldsson et al. (2007) simulated atmospheric flow over Iceland for the period September 1987 through June 2003 using version 3–5–3 of MM5 driven by initial and boundary data from the ECMWF. The simulated precipitation was compared with two types of indirect precipitation observations. Firstly, winter balance on two large outlet glaciers in SE-Iceland and on two large ice caps in central Iceland. Secondly, model output was used as input to the WaSiM hydrological model to calculate and compare the simulated runoff with observed runoff from six watersheds in Iceland for the water years 1987–2002. Model precipitation compared favourably with both types of validation data.

COMPARISON WITH GLACIOLOGICAL DATA

The spatial variability of the mass balance on large ice masses, such as Vatnajökull and Langjökull ice caps, can be mapped given data along several profiles extending over the elevation range of the ice caps. Mass balance has been observed on parts of Vatnajökull ice cap in SE-Iceland since 1991 (Björnsson et al., 1998) and from 1996 on Langjökull ice cap, central Iceland (Björnsson et al., 2002) (see location map on Fig. 2). Here, we use measurements of (accumulated) winter mass balance, expressed in terms of liquid water equivalents. Björnsson et al. (1998) estimated the uncertainty of the areal integrals of the mass balance to be a minimum of 15%. Due to surging of the Dyngjujökull glacier in 1998–2000, the uncertainty is considerably greater for this period and the following winter (Pálsson et al., 2002a). As yet unpublished data for the past few winters are from Björnsson and Pálsson[1]. The ice caps and typical locations of the mass balance stakes are depicted in Figure 2.

Mass balance on Hofsjökull ice cap has been observed at sites along profile HN (*cf.* Fig. 2) since 1987 and along profiles HSV and HSA since 1988 (Sigurðsson et al., 2004). In our model configuration the maximum elevation of the Hofsjökull ice cap is approximately 1540 metres, *i.e.* more than 250 metres lower than in reality. Hence, we use area-integrated data from an elevation range of approximately 1450–1650 metres along the three profiles HN, HSV and HSA (Jóhannesson et al., 2006). The winter balance on Hofsjökull has been modelled to estimate the amount of precipitation that falls as rain and ablation that may take place during the winter season. These estimates have been added to the measured winter balance to produce estimates of total precipitation at the measurement sites. This correction has not been carried out for Vatnajökull and Langjökull. The amount of liquid precipitation and winter ablation, therefore, has to be implicitly considered when comparison is made between precipitation simulated by MM5 and the glaciological measurements

[1]Helgi Björnsson and Finnur Pálsson, Institute of Earth Sciences and Science Institute, University of Iceland, personal communication.

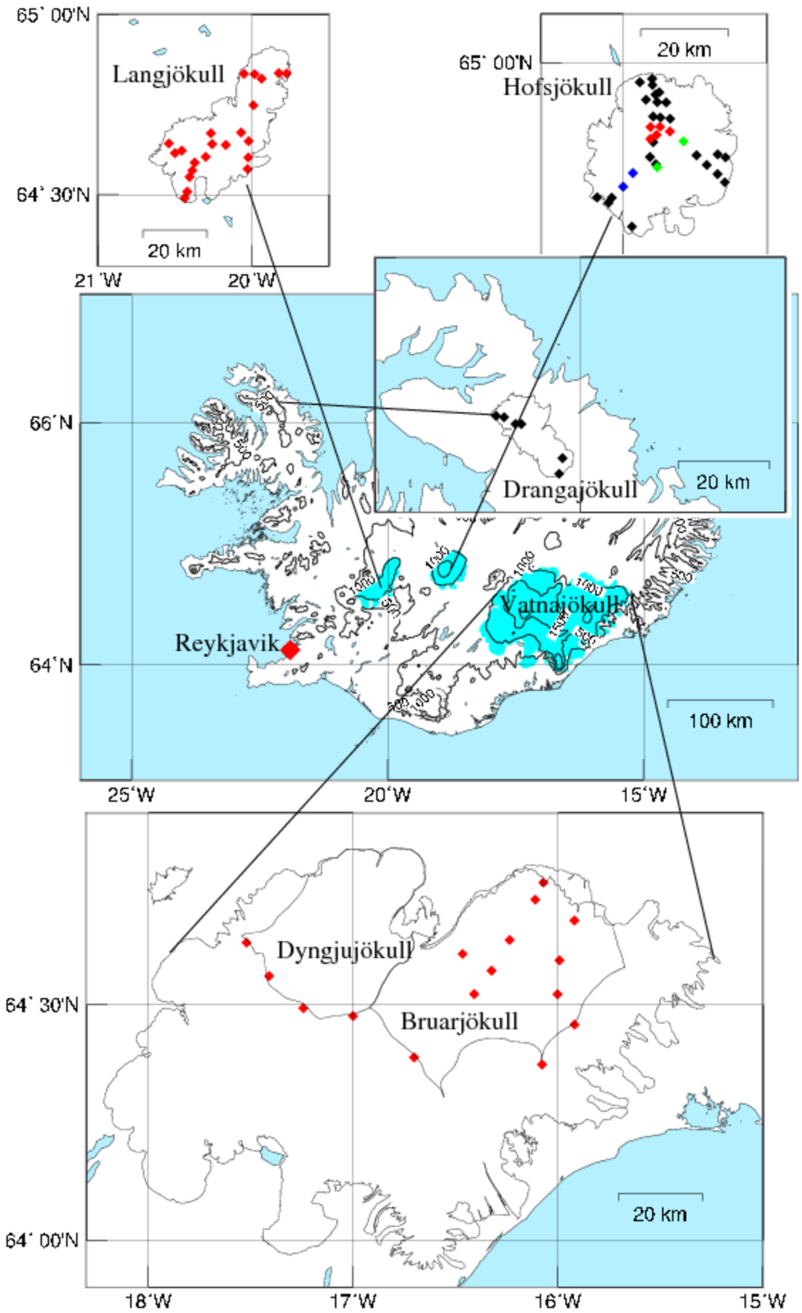

Figure 2: Overview of the six ice caps and glaciers used for validation purposes, dots indicate a typical location of observation sites. Red dots on Hofsjökull glacier are along profiles HN (N-part), blue dots along profile HSV (SW-part) and green dots along profile HSA (SE-part), observations at locations shown in black at Hofsjökull have not been used in this study. Drangajökull is split up in two regions, NW- and SE-part (*cf.* Table 1).

for Vatnajökull and Langjökull as discussed below.

The simulated winter precipitation at Hofsjökull ice cap is in good agreement with observations (*cf.* Fig. 3) over the northern part of the ice cap (HN,

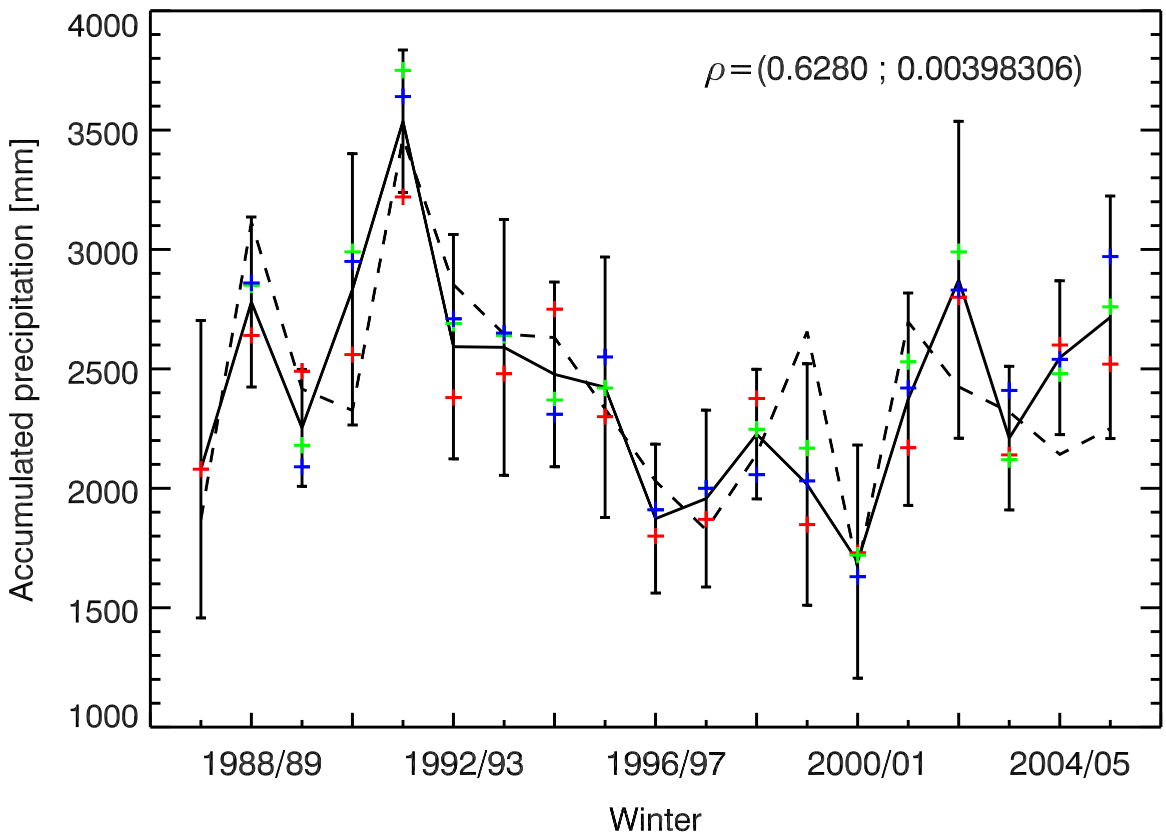

Figure 3: Estimated mean accumulated winter precipitation [mm] along profiles HN (N-part), HSA (SE-part) and HSV (SW-part) at altitudes between 1450 and 1650 metres (solid line, Jóhannesson et al., 2006). Dashed line represents simulated precipitation by MM5 (nine point average) at Hofsjökull ice cap. Red, green and blue crosses represent mean winter balance values at stakes along profiles HN, HSA and HSV respectively within the altitude interval 1440–1680 metres (*cf.* Fig. 2). Error bars indicate the standard deviation of the observations. Observed values from individual snow stakes are from Sigurðsson et al. (2004), Sigurðsson and Sigurðsson (1998) and Sigurðsson and Thorsteinsson (personal communication).

red dots, *cf.* Fig. 2), the SE-part (HSA, green dots, *cf.* Fig. 2) and the SW-part (HSV, blue dots, *cf.* Fig. 2). The solid line in Figure 2 shows the average of the observed winter precipitation, corrected to take liquid precipitation and/or winter ablation into account, at altitudes between 1450 and 1650 metres at locations HN, HSA and HSV. The dashed line represents precipitation simulated by MM5 (nine point average) at the location of the ice cap. The simulated precipitation is within one standard deviation of the average observed winter precipitation within this altitude range for sixteen out of nineteen winters during the period (1987–2006). The Spearman's rank correlation is 0.63 with a significance value of 0.004 and the RMS error is 49.

Areal integrals of winter balance over the Vatnajökull ice cap as a whole (8100 km^2), the Dyngjujökull (1040 km^2) and Brúarjökull (1695 km^2) outlet glaciers on the north side of the ice cap, and the Langjökull ice cap (925 km^2) are compared with simulated wintertime precipitation by the MM5 model in Figure 4. The winter balance is not corrected for to take liquid precipitation

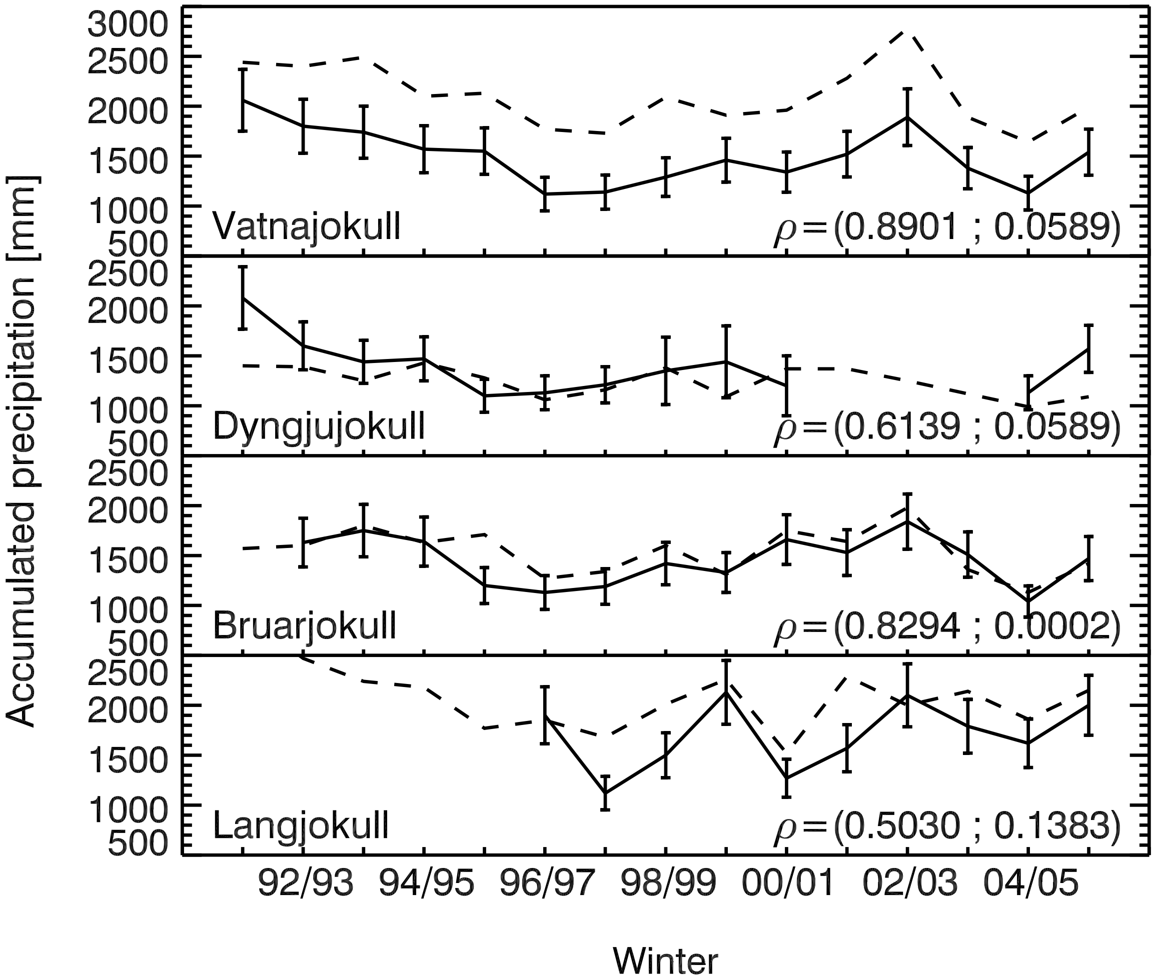

Figure 4: Observed accumulated winter balance (solid) and precipitation simulated by MM5 (dashed) for Vatnajökull ice cap as a whole (top), Dyngjujökull (second from top) and Brúarjökull (second from bottom) outlet glaciers and Langjökull ice cap (bottom). Error bars indicate 15% uncertainty of the observations, except for 1998–2001 at Dyngjujökull where it is 25%. Glaciological data for Vatnajökull, Dyngjujökull and Brúarjökull are from Björnsson et al. (1998, 2002) and Pálsson et al. (2002a,b, 2004b,c,d) Data for Langjökull ice cap are from Björnsson et al. (2002) and Pálsson et al. (2004a). As yet unpublished data for the past few winters are from Björnsson and Pálsson.

and/or winter ablation into account as mentioned above. The model shows least skill on Langjökull ice cap ($\rho = 0.50$; 0.14) where it has an RMS error equal to

372, and the greatest skill on Brúarjökull ($\rho = 0.83$; 0.0002) where the RMS error is equal to 171. The correlation for Dyngjujökull is 0.61 with a significance value of 0.06 and the RMS error is equal to 286. The simulated precipitation is within estimated observational error-margins for 10 out of 12 winters for Dyngjujökull, 13 out of 14 for Brúarjökull and 5 out of 10 for Langjökull ice cap. The correlation for Vatnajökull ice cap is 0.89, with a significant value of 0.06 and the RMS error is equal to 388. The relative importance of liquid precipitation and/or winter ablation is greatest for Vatnajökull as a whole because the southern margin of the ice cap reaches near sea level where rain may fall and ablation may take place at any time of the year. The north flowing outlet glaciers from Vatnajökull and Langjökull ice cap do not reach as far down so this problem is less important there. This is presumably the reason why the simulated winter precipitation is approximately 500 mm more than the observed winter balance for the Vatnajökull ice cap as a whole.

Mass-balance measurements at Drangajökull ice cap in NW-Iceland have only been carried out since 2004. Table 1 shows a comparison between simulated and observed winter balance for the mass-balance years 2004–2005 and 2005–2006 (Oddur Sigurðsson, personal communication). The model does not appear to capture the strong observed NW–SE precipitation gradient. The single grid cell values for the SE-part are very close to the observed values but they are too high for the NW-part. The area-averaged values from MM5 are, however, close to mean observed values for the NW-region of the ice cap but too low for the SE-part.

Table 1: Accumulated winter balance and simulated wintertime precipitation at Drangajökull, NW-Iceland (*cf.* Fig. 2). Observed winter balance is taken as the mean of stakes above 400 metre altitude in the northwestern (NW) part of the ice cap and in the southeastern (SE) part. Simulated precipitation is both taken as a nine point mean value (lower values) for the nearest grid cells as well as the nearest grid cell value (higher values).

Winter	**NW_{Obs} [mm]**	**NW_{MM5} [mm]**	**SE_{Obs} [mm]**	**SE_{MM5} [mm]**
2004/05	1797 (3 pts.)	2090/2554	2675 (2 pts.)	2072/2603
2005/06	1833 (3 pts.)	2105/2524	2815 (2 pts.)	2127/2604

CONCLUSIONS

In general, the MM5 model results compare favourably with the observed winter balance, in particular for Hofsjökull, where corrections to take liquid precipitation and/or winter ablation into account have been made, and for the comparatively high altitude outlet glaciers Dyngjujökull and Brúarjökull, where such corrections are relatively unimportant. More extensive comparison of simulated precipitation with glaciological observations needs to be made with corrected mass balance data from all the ice caps.

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DYNAMICAL DOWNSCALING OF PRECIPITATION – PART II: COMPARISON WITH RAIN GAUGE AND HYDROLOGICAL DATA

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ABSTRACT

Atmospheric flow over Iceland has been simulated for the period January 1961 to July 2006, using the mesoscale MM5 model driven by initial and boundary data from the ECMWF. Firstly, a systematic comparison of results to observed precipitation has been carried out. Undercatchment of solid precipitation is dealt with by looking only at days when precipitation is presumably liquid or by considering the occurrence and non-occurrence of precipitation. Away from non-resolved orography, the long term means (months, years) of observed and simulated precipitation are often in reasonable agreement. This is partly due to a compensation of the errors on a shorter timescale (days). Probability of false alarms (model predicts precipitation, but none is observed) is highest in N-Iceland, particularly during winter. The probability of missing precipitation events (precipitation observed but none is predicted by the model) is highest in the summer inland in N-Iceland. Secondly, model output is used as input to the WaSiM hydrological model to calculate and compare the runoff with observed runoff from six watersheds in Iceland. It is found that model results compare favourably with observations.

INTRODUCTION

The 6–hourly ERA40 re-analysis of the ECMWF have been dynamically downscaled for the period 1961–2006 using the numerical model MM5 (Grell et al., 1995). The modelling approach is described in detail in Rögnvaldsson et al. (2007, 2008).

Climatological downscaling of precipitation is of use for hydrological purposes. The MM5 model, using a similar setup as used in this study, is in operational use in Iceland for production of short to medium range weather forecasts. Although a hydrologist and a weather forecaster would both like to be able to predict precipitation, their interests lie on different timescales.

In this paper we use output from the MM5 model as input to the WaSiM hydrological model (Jasper et al., 2002) for the same six watersheds as used for validation purposes by Rögnvaldsson et al. (2007) and compare the simulated discharge with observed discharge. We also evaluate the quality of the simulations by comparing them to rain gauge measurements. This can be done by comparing long term means (months, years) of simulated and observed precipitation. Such a comparison would be of use to a hydrologist but of somewhat limited value for a forecaster. We therefore set out to making comparisons that would assess strong and weak points of the simulations to aid forecasters. We want to know how the errors in the simulated precipitation relate to other meteorological factors and if the performance depends on the temporal resolution of the data and geographical location. The work should shed a light on which aspects need improving. Increased understanding of the limitations of the simulations on a short timescale will also be beneficial for their use in hydrological purposes.

This paper begins with a short description of rain gauge data used in this study and how simulated precipitation compares to observations. Following is a comparison of modelled discharge to observed discharge and concluding remarks.

RAIN GAUGE DATA

The dynamic downscaling of ECMWF data, using version 3–7–3 of the MM5 model, has been compared to precipitation observations from synoptic stations for the sub-period 1987 2003. Precipitation is measured twice per day on the chosen synoptic stations, at 09 and 18 UTC. The MM5 output was saved every 6 hours, at 00, 06, 12 and 18. The shortest comparison period is therefore 24 hours (from 18 to 18). That period will from now on be referred to as an "event" in this paper.

The model output from a grid point can be considered as an area averaged precipitation over an area of 64 km^2. Therefore we do not expect the simulations to agree with measurements in areas with topography that is not resolved by the model. When comparing simulated and observed precipitation we must also bear in mind the general problems of precipitation observations. The most significant of these is the large undercatchment of solid precipitation in cold and windy climate, as in Iceland (Førland et al., 1996). Undercatchment of solid precipitation is dealt with by looking only at days when precipitation is presumably liquid (summer or temperature criteria) or by considering the occurrence and non-occurrence of precipitation.

COMPARISON WITH OBSERVED PRECIPITATION

Figure 1 shows the relative error of the simulations, (mm5-obs) / obs, for the summer months June, July and August (JJA). It can be seen that the model

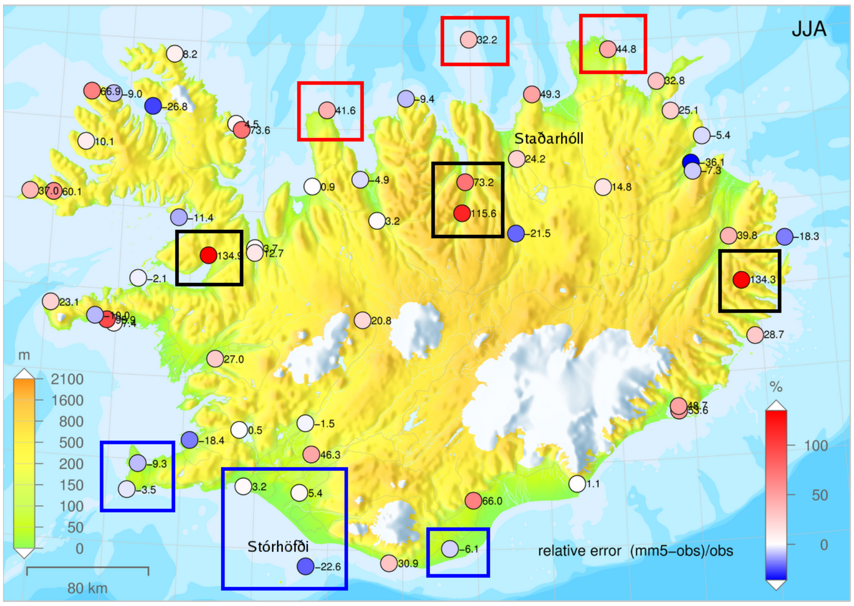

Figure 1: A topographic map of Iceland showing relative difference between simulated and observed accumulated precipitation, (mm5-obs) / obs. Each coloured circle corresponds to a synoptic weather station. Station names are included at the stations referred to in this paper. The colour of the circle denotes the relative error in the simulations (colourbar to the right) for the summer months June, July and August (JJA). The blue boxes enclose a few stations on flat land in S-Iceland where the observations and simulations are in reasonable agreement. The red boxes draw attention to stations in N-Iceland where the model overestimates precipitation, despite these stations being on flat land. Stations that have huge overestimation, which is almost certainly due to non-resolved orography, are enclosed in black boxes.

behaves differently in N- and S-Iceland for stations in flat land (minimal effect of non-resolved orography). For stations on flat land in the South, the simulations and observations are in an overall reasonable agreement (see stations in blue boxes in Fig. 1). The model does however underestimate precipitation in flows from the SE (not shown). The model overestimates the precipitation for flat land stations in the North (red boxes in Fig. 1). This is particularly true in northerly flow. For stations situated in orography that is obviously not resolved by the model (black boxes in Fig. 1), the somewhat expected result of huge relative errors is clearly visible.

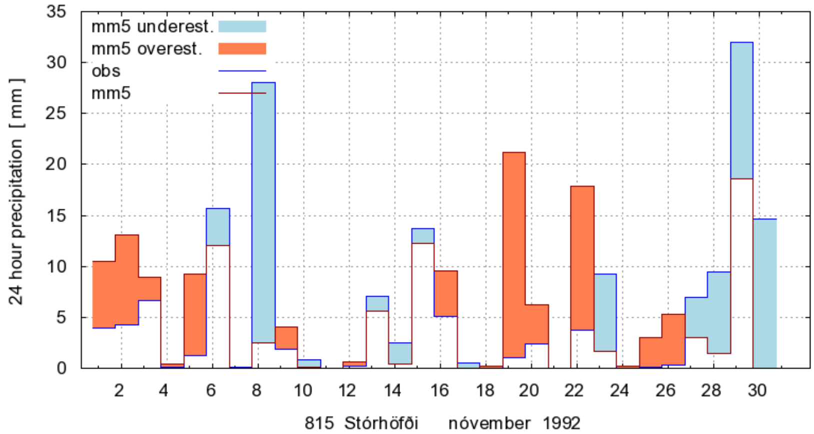

Figure 2: Data from Stórhöfði, S-Iceland, accumulated 24 hour precipitation [mm] (observed and simulated) for November 1992. Blue colour denotes the amount of mm5 underestimation and red denotes the mm5 overestimation.

The 24 hour precipitation amounts (observed and simulated) for November 1992 at Stórhöfði, S-Iceland, is shown in Fig. 2. The sums of observed and simulated precipitation for this month are almost identical. It is however clear that the agreement of the monthly sums is in large part due to compensation of the errors on a daily timescale. We define a "false alarm" event as a period of 24 hours (from 18 to 18) where there is some precipitation in the simulations ($r_{mm5} > 0.1$) but the observations are dry ($r_{obs} \leq 0.1$). Figure 3 shows the percentage of events that fall into the false alarm category on each of the stations during the winter months December, January and February (DJF). Comparison with maps from the other seasons (not shown) reveals that there is increased probability of false alarms in winter, most notably for inland areas in N-Iceland. In Fig. 4 all false alarm events at Staðarhóll have been categorized according to wind direction. We see that most of the precipitation during false alarm events is associated with southerly winds.

A "missing" event is defined as a 24 hour period where the simulations are dry ($r_{mm5} \leq 0.1$) but the observations are wet ($r_{obs} > 0.1$). Figure 5 shows the percentage of missing events during the summer months (JJA) at each of the observation stations. There is higher probability of missing events during summer than in winter (map not shown). In Fig. 6 the precipitation during missing

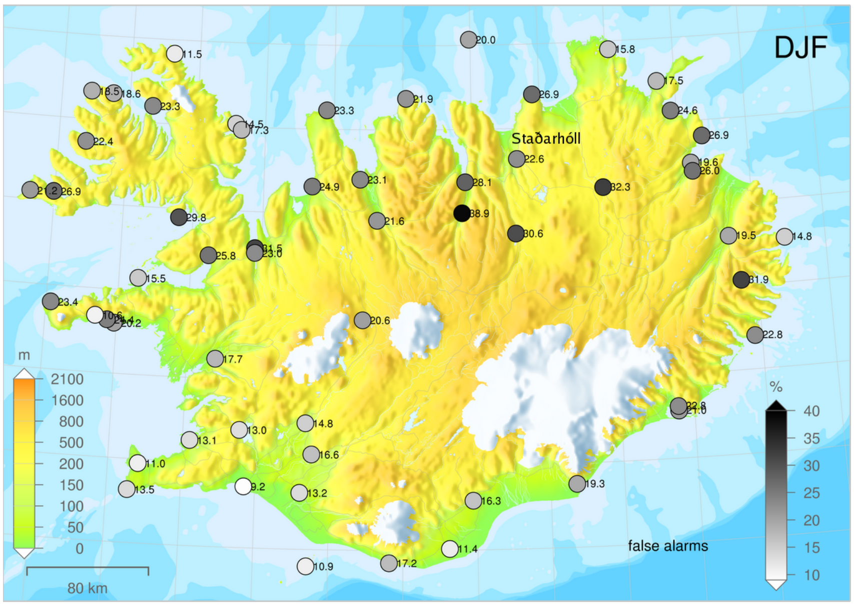

Figure 3: Ratio [%] of "false alarms" (mm5 wet, obs dry) during winter (DJF).

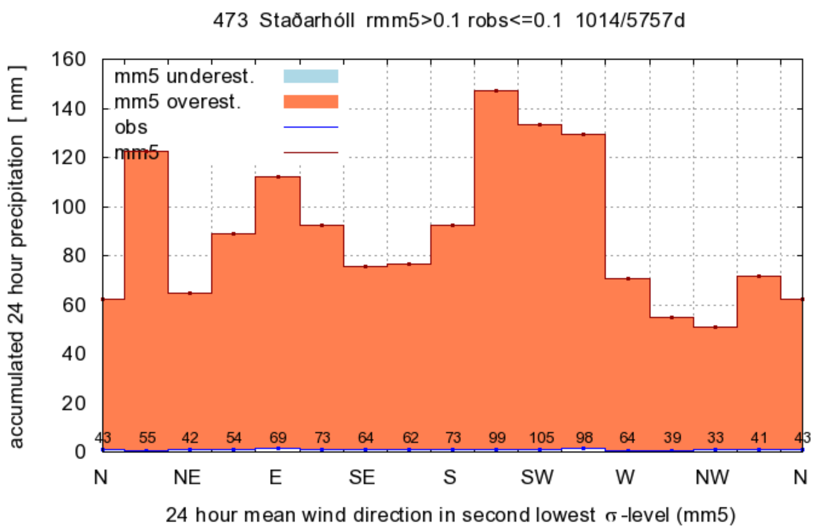

Figure 4: All "false alarm" events from Staðarhóll, NE-Iceland. The horizontal axis shows bins for 16 wind directions. The vertical axis shows the accumulated precipitation in each bin.

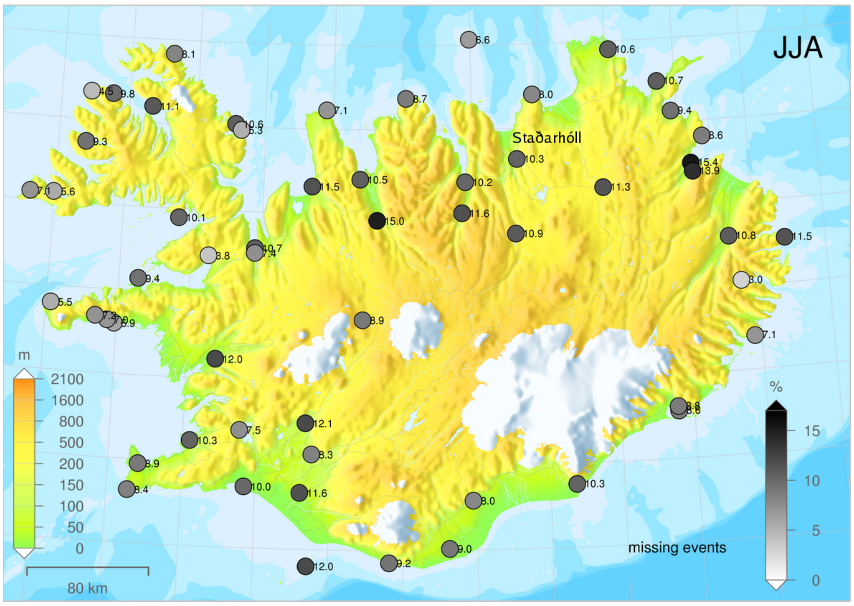

Figure 5: Ratio [%] of "missing" events (mm5 dry, obs wet) during summer (JJA).

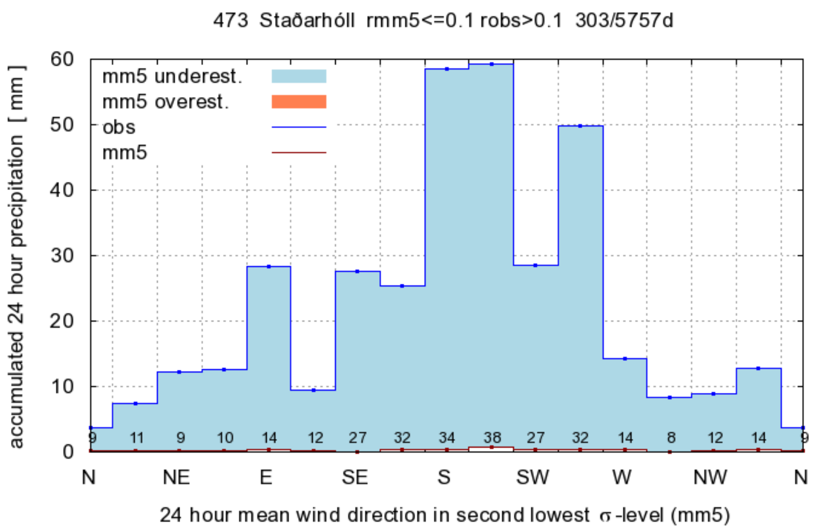

Figure 6: All "missing" events from Staðarhóll, NE-Iceland. Horizontal axis shows bins for 16 wind directions. The vertical axis shows precipitation sum in each bin.

events (only observed precipitation) at Staðarhóll has been grouped into bins of different wind direction and the precipitation in each bin added up. Again we see that southerly winds (lee side) are the main culprit.

COMPARISON WITH HYDROLOGICAL DATA

Several authors have used runoff measurements for validation of precipitation simulated by atmospheric models. Benoit et al. (2000) reported some of the advantages of using one-way coupling of atmospheric and hydrological models, calibrated with observed discharge data, for validation of precipitation calculated by the atmospheric models. They concluded that stream flow records give a better estimate of the precipitation that has fallen over a region than point precipitation measurements, and even though there were uncertainties related to their hydrological model (WATFLOOD), it was sufficiently sensitive to help improve atmospheric models. Jasper and Kaufmann (2003) compared results from WaSiM watershed models that were on one hand driven by meteorological observations and on the other hand driven by data from atmospheric models. They concluded that the hydrological model was sufficiently sensitive to provide substantial information for the validation of atmospheric models.

Jónsdóttir (2008) used the latest output from version 3–7–3 of the MM5 model as input to the WaSiM model for the period 1961–1990 to create a runoff map of Iceland. The difference between measured and modelled discharge was in general found to be less than 5% although larger discrepancies were observed (see Fig. 7). For a full list of stations we refer to Table 2 in Jónsdóttir (2008)). The WaSiM model was not run with a groundwater module. Instead precipitation simulated by MM5 was scaled in order to make the simulated water balance fit the measured water balance for individual watersheds, a detailed description can be found in Section 6 in Jóhannesson et al. (2007) and Jónsdóttir (2008). Therefore, comparison of measured and simulated water balance cannot be directly used for validation of the model-generated precipitation. According to the non-scaled MM5 output for the period 1961–1990, mean precipitation for the whole of Iceland was 1790 mm y^{-1}. After scaling the precipitation, this value was reduced to 1750 mm y^{-1}, *i.e.* by approximately 2%. This difference can, to some extent, be explained by the fact that precipitation falls on porous post-glacial lava in some areas and flows through groundwater aquifers to the ocean without participating in surface runoff. Earlier research (Tómasson, 1982) have estimated this flow to be on the order of 33–62 mm y^{-1}. This comparison of total accumulated scaled and non-scaled precipitation indicates that MM5 produces comparatively unbiased precipitation estimates when integrated over the

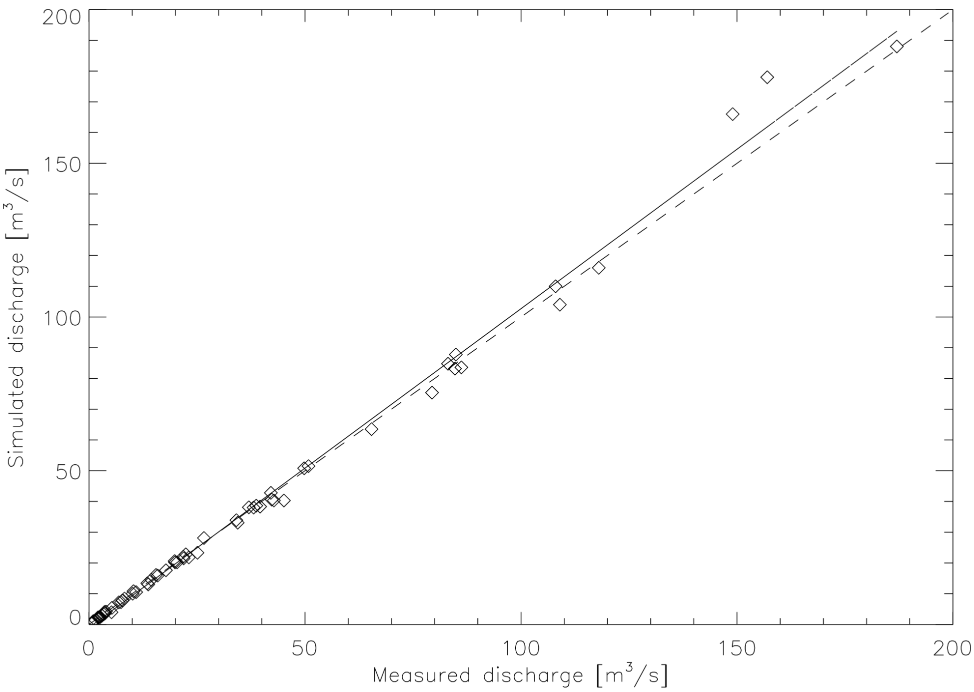

Figure 7: Measured and simulated (WaSiM / MM5) mean discharge [m^3s^{-1}] at the watershed gauges. Dashed line indicates a perfect fit, solid line represents the linear best fit between the measured and simulated discharge.

whole of Iceland.

Table 1 compares observed and modelled discharge from six watersheds that are not much affected by groundwater flow (the same discharge stations as used for validation of an earlier MM5 model version by Rögnvaldsson et al., 2007; *cf.* Table 1 and Fig. 2). Here, un-scaled precipitation is used in the hydrological modelling in order to obtain an independent validation of the precipitation generated by MM5. For four out of six watersheds, the difference in the water balance is reduced when the newer version of the MM5 model is used compared with the results obtained with the earlier model version. The relative difference between the simulated and observed water balance is in the range −8 to 13%, with four of the six values in the range −4 to 5%, indicating a satisfactory performance of the model.

CONCLUSIONS

The numerical model MM5, run at a horizontal resolution of 8 km, has been used to downscale over Iceland the 6-hourly analysis of the ECMWF over a period of 46 years. A systematic comparison with observed precipitation for the sub-period 1987–2003 has been presented as well as comparison of simulated discharge with observed discharge. The main results are:

Table 1: Comparison of observed and simulated discharge [m^3s^{-1}] at six discharge stations using unscaled modelled precipitation from versions 3-5 and 3-7 of the MM5 model. Note that the simulation periods are not the same for the two model versions. Hence, the measured discharge can differ somewhat between the columns corresponding to the two versions. The discharge stations are, respectively; Vatnsdalsá river, Norðurá river, Fossá í Berufirði river, Hvalá river, Fnjóská river and Hamarsá river. The simulation periods are, respectively; 1963–2001, 1971–2001, 1963–2001, 1976–2001, 1976–2001 and 1991–2004.

Station #	MM5 V3-5			MM5 V3-7		
	Q_{meas}	Q_{calc}	Difference	Q_{meas}	Q_{calc}	Difference
45	12.3	13.4	8.9%	10.3	10.8	5.0%
128	29.4	32.2	9.7%	22.4	25.3	13.0%
148	9.1	10.4	14.3%	8.2	7.9	−4.0%
198	26.8	25.4	−5.2%	15.5	15.3	−1.0%
200	48.4	53.9	11.4%	39.6	40.3	2.0%
265	19.6	20.8	6.1%	19.9	18.4	−8.0%

- Away from non-resolved orography, long term (months, years) sums of simulated precipitation are quite correct in the south but too high in the north. This is partly due to compensating errors on a smaller time scale (days).
- Probability of false alarms (model predicts precipitation, but none is observed) is highest in N-Iceland, particularly during winter.
- Probability of missing precipitation events is highest in the summer inland and on the lee side of Iceland in southerly flows.
- Precipitation is underestimated in SE flows in SW-Iceland but precipitation is overestimated in northerly flows in N-Iceland. This cannot only be explained by non-resolved orography.
- Simulated discharge compares favourably with observed discharge for the majority of observation sites, indicating a satisfactory performance of the model.

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A SCALING METHOD FOR APPLYING RCM SIMULATIONS TO CLIMATE CHANGE IMPACT STUDIES IN HYDROLOGY

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ABSTRACT

Climate change may have significant influence on hydrology and corresponding water management, such as large-scale flooding, reservoir regulation and others. In analysing projected future conditions, it is therefore important to use the most appropriate climate change signals for regional and local hydrological studies. In recent years, Regional Climate Models (RCMs) have developed rapidly to improve representation of local climate situations. However, systematic biases still constrain direct use of RCM outputs for impact studies. A scaling method has therefore been developed to adjust meteorological variables from RCMs to better reflect present-day conditions during representative control periods. The method adjusts biases identified for daily precipitation and temperature. Statistical properties, such as mean and standard deviation of daily precipitation and temperature, are much improved compared to direct RCM output. More realistic frequency of rainy days and distribution of daily precipitation and temperature are generated. The scaled precipitation and temperature were used as inputs to the HBV rainfall-runoff model for several river basins in Sweden. The resulting runoff showed good consistency with historical records and is thought to be better suited for climate change impact studies.

INTRODUCTION

Increases of radiatively active gases in the atmosphere will most likely lead to a change in the climate, and this will in turn most likely lead to an intensification of the hydrological cycle on global and regional scale (Huntington, 2006). The best tools to estimate the character of a climate change are Global Circulation Models (GCMs) that model the interaction between atmosphere, ocean and biosphere on a global scale. Output from GCMs cannot be directly used in impact models, such as rainfall-runoff models, because of biases in modelled variables (Murphy et al, 2004). There is therefore a need to downscale the output from GCMs to regional and local scales.

Regional Climate Models (RCMs), with boundary conditions from an overlying GCM, transfer the climate change signal to regional scale and incorporates many of the local conditions, such as orography, that improves representation of many important variables. There are, however, still problems with biases when output from an RCM is compared with observed data. Much of the bias originates from the GCM, but the parameterisation in the RCM can also produce model biases (Graham et al, 2007). As a result, RCM scenario simulations over Scandinavia tend to have too many rainy days and too few days with extreme precipitation. This unfavourable 'drizzle' feature combined with biases in daily temperature results in unrealistic climate conditions at catchment scales, and the magnitude of important events such as floods and droughts are often underestimated.

This problem can be addressed by perturbing an observed data series with a projected future change, also known as the delta change method, instead of using RCM output directly. However, this approach does not accommodate possible changes in climate variability (Lenderink et al, 2007). Another method to correct the problem is to scale the RCM output on the basis of a comparison with observed weather variables over a control period, also referred to as bias correction (Leander and Buishand, 2007). The derived correction factors can then be applied to RCM output representing future climate to create a more realistic assessment of the change in hydrological responses. The advantage of this method is that it is relatively easy to implement over large areas, provided that good observed data is available. The caveat of the method is that the scaling might disrupt important climate change signals. In this paper, a new scaling method for RCM output is developed and tested for a catchment in central Sweden.

DATA

In this paper, the river basin of Torpshammar was selected to validate the model performance. The Torpshammar basin is situated in central Sweden (Fig. 1) and is a part of River Gimån, the largest tributary to River Ljungan. The basin area is approximately 4 300 km^2 and the mean altitude is about 350 m.a.s.l. Eighty percent of the basin area is forested and the rest is equally divided between open land and lake area. The annual mean precipitation is around 560 mm and the average temperature is +1.7 °C (1961-1990). There are 6 sub-basins in the main set-up of the hydrological HBV model (Lindström et al., 1997) for Torpshammar, but for simplicity only one sub-basin was selected to show the results.

A regional climate scenario for Europe was provided by the Rossby Centre Atmospheric model, RCA3 (Kjellström et al., 2006). Boundary conditions for the RCA3 model were in turn provided by the ECHAM5 GCM (Roeckner et al., 2006). Assumptions for future green house gas concentrations came from the emission scenario SRES A1B (Nakićenović et al., 2000). The notation for the run is R3E5A1B and it is available for the time period 1961-2100.

A gridded dataset (4x4 km^2) of observed temperature and precipitation has been used as the baseline in the scaling procedure. The dataset was calculated from a network of observation stations in Sweden using optimal interpolation (Johansson, 2000) and is available from 1961 to present day. Observations in the database have been adjusted according to the effects from topography, wind direction and wind speed (Johansson and Chen, 2003; 2005).

Figure 1. The river basin of Torpshammar.

METHODOLOGY

Scaling method for precipitation

A key assumption to apply the scaling approach is that the climate statistics over a sufficiently long time period from the climate model simulation is comparable to observed climate statistics. In this study we have assumed that 30 years are sufficient and used the time period 1961-1990 from both observations (reference) and from the RCM (control).

The method to scale RCM precipitation was carried out in two steps. Firstly, a cut-off value for precipitation was used to reduce the number of rainy days in the total period, in light of the spurious drizzle generated by the RCA3 model. The second step was to transform the modelled precipitation to realistic amounts, in particular on days with heavy rainfall. This process was performed separately for each sub-basin in order to preserve the spatial variability of the corrected output.

For individual sub-basins, observed and RCA3-generated precipitation time series were compared to identify the cut-off threshold value for the RCA3 output. The wet days generated by RCA3 with amounts larger than the threshold were considered as rainy days, and all other days as dry.

There are a number of statistical distributions that can be used to describe the probability distribution of precipitation intensities, for instance exponential distribution (Todorovic and Woolhiser, 1975). In this study the gamma distribution was selected, since it can represent the typically asymmetrical and positively skewed distributed daily precipitation (Wilks, 1995). Its density distribution is expressed as:

$$f(x) = \frac{(x/\beta)^{\alpha-1} \cdot \exp(-x/\beta)}{\beta \cdot \Gamma(x)} \qquad x,\ \alpha,\ \beta \succ 0 \tag{1}$$

where α is the shape parameter, β the scale parameter and $\Gamma(x)$ is the inverse gamma function. The distribution parameters were estimated using maximum likelihood estimator (MLE).

Precipitation events with a low volume are much more frequent than events with a high volume, and therefore the small events tend to dominate in the parameter estimation. In order to capture the main properties of normal rainfall events as well as extreme rainfall events, the precipitation time series were divided into two partitions separated by the 95th percentile value. The resulting distribution is named double gamma distribution that is distinguished from single gamma distribution using all the events to fit one distribution. Two sets of parameters – α, β and α_{95}, β_{95} – were respectively estimated from observations and RCA3 output in the control period and these parameter sets were in turn used to correct the RCA3 future scenario output (Eq. 2).

$$\begin{cases} P_{Scaled} = F^{-1}(\alpha_{Obs}, \beta_{Obs}, F^{-1}(x, \alpha_{CTL}, \beta_{CTL})) & \text{if } x \prec 95^{th} \text{ percentile value} \\ P_{Scaled} = F^{-1}(\alpha_{Obs,95}, \beta_{Obs,95}, F^{-1}(x, \alpha CTL,95, \beta_{CTL,95})) & \text{if } x \geq 95^{th} \text{ percentile value} \end{cases} \tag{2}$$

where α_{Obs} and β_{Obs}, $\alpha_{Obs},_{95}$ and $\beta_{Obs},_{95}$ are parameters from observations; α_{CTL} and β_{CTL}, $\alpha_{CTL},_{95}$ and $\beta_{CTL},_{95}$ are parameters from the RCA3 output in the control period; x is the daily precipitation from the RCA3 output. F stands for the probability distribution of gamma distribution.

The transform parameters were optimised separately for the four seasons DJF (December, January, February), MAM (March, April, May), JJA (June, July, August) and SON (September, October, November). Note that season classification could in future work be varied depending on regional climatic characteristics.

Scaling method for temperature

Temperature was scaled using the statistics on a daily scale. Since there is a relationship between temperature and rain, the scaling was done conditionally on occurrence of precipitation on the target day. Long-term averages of temperature, μ, and its standard deviation, σ, were calculated for each day. They were described with Fourier series (Eqs. 3 and 4):

$$\mu\left(t^*_{Dry/Wet}\right) = \frac{a_{0,Dry/Wet}}{2} + \sum_{k=1}^{K}\left(a_{k,Dry/Wet}\cdot\cos(kwt^*) + b_{k,Dry/Wet}\cdot\sin(kwt^*)\right) \tag{3}$$

$$\sigma\left(t^*_{Dry/Wet}\right) = \frac{c_{0,Dry/Wet}}{2} + \sum_{k=1}^{K}\left(c_{k,Dry/Wet}\cdot\cos(kwt^*) + d_{k,Dry/Wet}\cdot\sin(kwt^*)\right) \tag{4}$$

where, a_0, a_k, b_k, c_0, c_k and d_k are the Fourier coefficients, k stands for the n^{th} harmonic used for describing the annual cycle of daily temperature. By using the Fourier series, the annual cycle can be described with just a few parameters. The parameters were calculated from dry and wet days of the observations and model simulations, respectively, in the control period to specify the scaling factors. Thus, the bias in daily temperature can be corrected. The calculated parameters can be further applied to future climate (Eq. 5):

$$T_{scaled,Dry/Wet} = G^{-1}(\sigma_{Obs,Dry/Wet}, \mu_{Obs,Dry/Wet}, G^{-1}(y, \sigma_{CTL,Dry/Wet}, \mu_{CTL,Dry/Wet})) \tag{5}$$

Where, $\mu_{Obs,Dry/Wet}$ and $\sigma_{Obs,Dry/Wet}$ are parameters from observations; $\mu_{CTL,Dry/Wet}$ and $\sigma_{CTL,Dry/Wet}$ are parameters from the RCA3 output in the control period; y is the daily temperature from the RCA3 output. G represents the probability distribution of normal distribution.

RESULTS

Scaling improved the RCA3 distribution of precipitation intensities (Figs. 2 and 3). Both single gamma and double gamma distributions were tested on observations and RCA3 output, and in the climate impact extension to the future scenario precipitation. The method using double gamma distributions was superior than using a single gamma transformation in terms of adjusting the daily precipitation (Fig 2.)

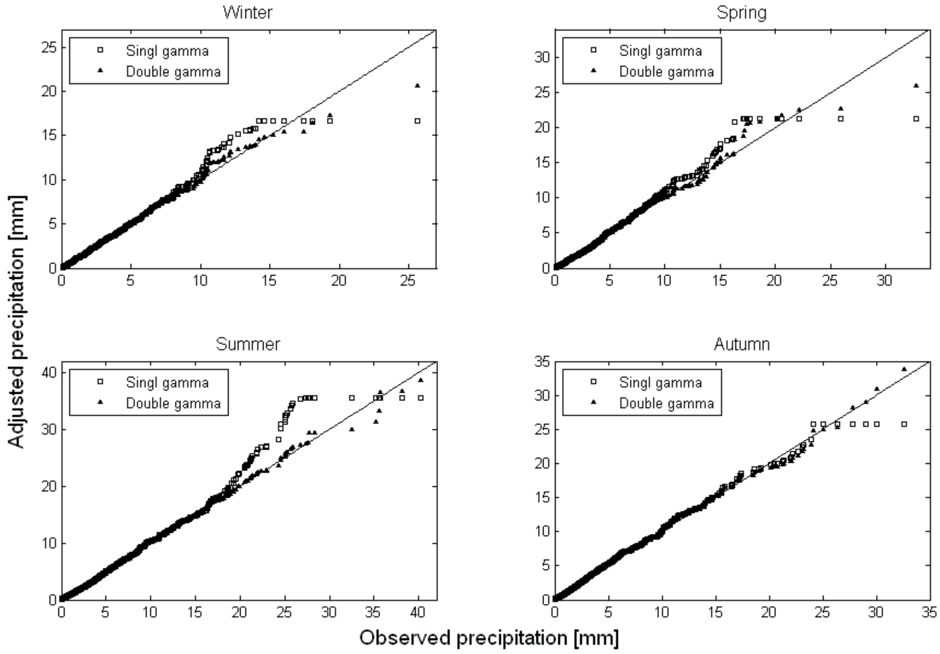

Figure 2. Comparison of observed and scaled RCA3 precipitation in Torpshammar for the control period of the R3E5A1B simulation.

The large differences in rainfall frequency and amount between the observation and simulation are reduced considerably for each season. The double gamma distribution was able to reproduce large rainfall events well under extreme weather conditions. Daily average temperature is scaled for each sub-basin. After scaling, both mean and variability of daily temperature were improved.

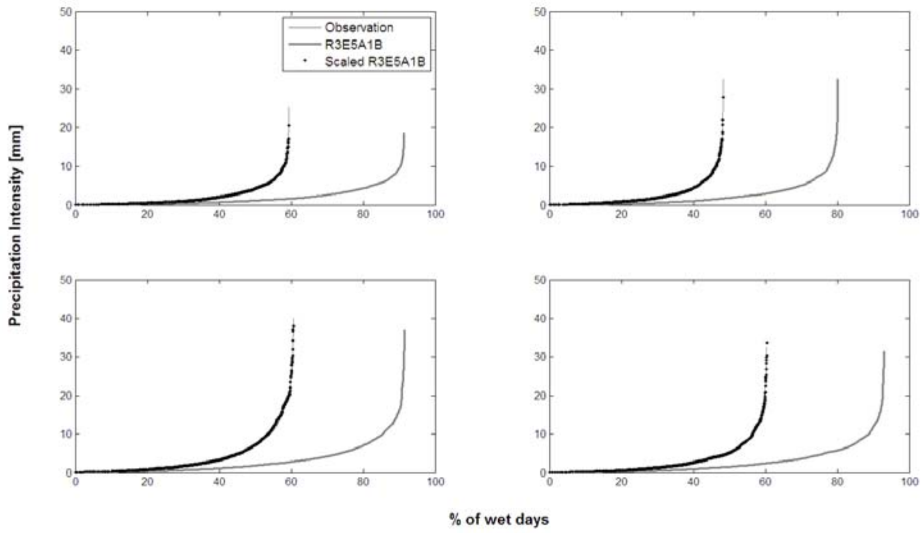

Figure 3. Distribution of precipitation intensity from observations, R3E5A1B simulation and scaled R3E5A1B simulation for each season in Torpshammar.

For daily temperature the method improved the modelled output considerably (Fig 4). The largest improvements are shown in correcting the warm bias in the winter and the cold bias in the summer.

After scaling, adjusted daily precipitation and temperature were used as inputs to drive the HBV rainfall-runoff model for investigation of hydrological response. Compared to the runoff generated directly from RCA3 simulation, the one generated with scaled P and T shows much better agreement with observations (Fig 5). The peak runoff in spring was correct in terms of both volume and timing and the simulated snow cover is in good agreement with that calculated from the observation.

DISCUSSION AND CONCLUSIONS

Many of the catchments in northern Sweden have their maximum peak flow during the spring flood, thus biases in precipitation and temperature create potentially large errors in the accumulation of snow and the subsequent spring flood. On these scales, the bias in the GCM/RCM outputs makes it necessary to perform corrections before they can be used in impact studies. The scaling of temperature and precipitation performed in this study removed most of the bias in the RCA3 output, making the modelled precipitation and temperature series more useful for hydrological impact studies in a future climate. A caveat with this approach is that a further step in the modelling chain is added, which might increase uncertainty.

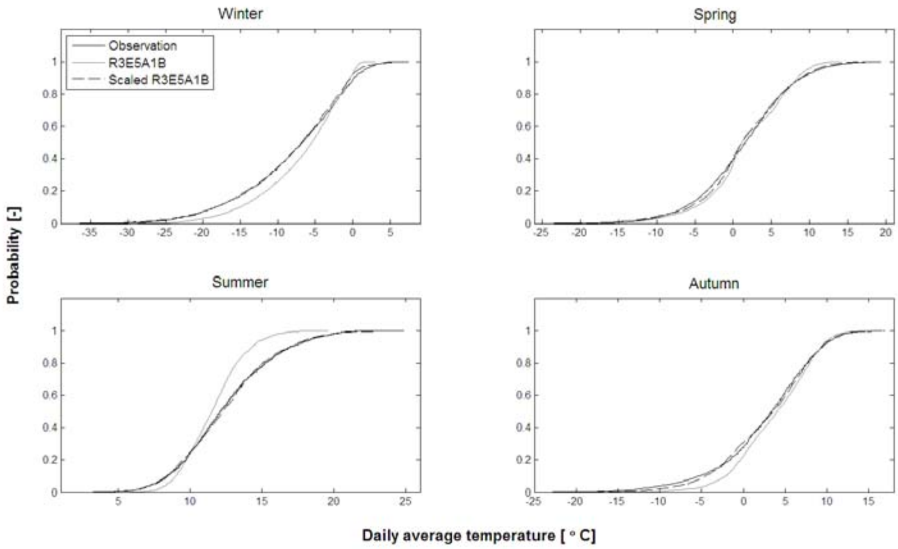

Figure 4. Distribution of daily temperature from observations, R3E5A1B simulation and scaled R3E5A1B simulation for each season in the Torpshammar catchment.

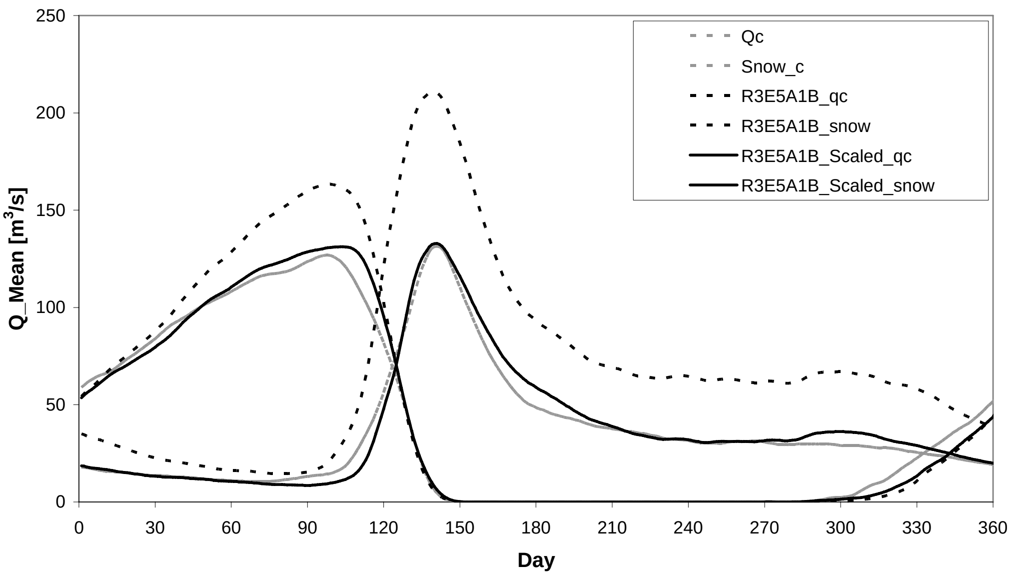

Figure 5. Mean daily runoff (Qc) and snow cover (snow) in Torpshammar calculated using precipitation and temperature from observations, R3E5A1B simulation and scaled R3E5A1B simulation.

Using a double gamma distribution was favoured over using a single gamma distribution (Fig. 2). Too much emphasis is put on low precipitation values if only one gamma distribution is used and the extreme events are not well captured. The inclusion of a second gamma distribution for precipitation values above the 95^{th} percentile solved this problem.

Correcting only the bias in the precipitation distribution is not sufficient to model the runoff correctly, especially if maximum flows occur during the spring flood. The volume during the spring flood is mainly determined by accumulated snow. The timing and volume of the maximum flow is highly dependent on a correct description of temperature during winter and spring. Good model representation of the spring flood is important in northern Swedish rivers to estimate potential flooding and to efficiently plan production of electricity in the hydropower plants. A correct description of temperature is also important to reproduce low flows caused by droughts in catchments in the southern part of Sweden.

GCM/RCM simulations are our best tools to assess how a climate change may affect the water resources and possible increased risk for natural hazards such as flooding and drought. However, hydrological impact models are dependent on good representation of the driving variables P and T, and scaling of RCM output effectively removed the biases in these variables.

The application of the here developed scaling procedure has been started for all of Sweden, to adjust simulations from different GCM/RCM combinations. This will include studies of the effect of different initial conditions in the GCM and representing different emission scenarios. The scaled ensemble will provide a good picture of the uncertainties involved and a sound basis for hydrological assessment.

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THE NET EFFECT OF SAMPLE VARIABILITY AND RATING CURVE IMPRECISION IN REGIONAL FLOOD FREQUENCY ANALYSIS

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ABSTRACT

This study examines the joint impact of sample variability and rating curve imprecision in regional instantaneous flood frequency analysis based on L-moments. A parametric bootstrap methodology is developed for this purpose, assuming (1) a power-law model for the stage-discharge measurements, (2) a generalised extreme value (GEV) model for the annual maximum discharges and (3) no intersite correlation. The bootstrap framework is applied to data from eight gauging stations located in the southwest of Norway. The application shows that rating curve imprecision can have a major impact on the accuracy of regionally-based T-year flood estimates. The coefficients of variations for 25- and 50-year flood return level estimates at two of the eight stations are used as case-studies. The main findings is that when more stations are added to the regional analysis, the effect of sample variability decreases rapidly before stagnating, whereas the uncertainty due to rating curve imprecision fluctuates around what appears to be an average value. The implications of these results are discussed.

INTRODUCTION

Consider the problem of preventing or decreasing the impacts of future floods. The first task is to make inferences about their estimated magnitudes. In addressing this problem one is faced with a number of uncertainty components, e.g. model uncertainty, the uncertainty associated with the estimation of the parameters of the chosen model due to limited data and the uncertainty caused by inaccurate measurement of the available data. This study highlights the last two components.

Numerous studies have explored the benefit of using additional data from other gauging stations in the face of sample variability (SV). It is now commonly accepted in hydrology that a regional flood frequency analysis provides a more rational and accurate method than the at-site approach. However, uncertainty in streamflow data, of which rating curve imprecision (RCI) is the greatest contributor, is not included in the basis for this conclusion. Typically, a rating curve is derived by fitting a power-law model to the available stage-discharge measurements. Assuming the power-law relationship is true, the uncertainty in this procedure can be summarised in three categories: (1) if the stage-discharge relationship is compound, uncertainty in segmentation, (2) variability due to discharge measurement inaccuracy and (3), if extrapolation is performed, uncertainty whether the fitted curve is valid in the unexplored area. This study investigates category (2).

The literature on sample variability in regional flood frequency analysis is extensive (see for example Kjeldsen and Jones, 2006, for details and references). In contrast, the impact of rating curve imprecision in the estimation of the T-year flood has received very little attention. Potter and Walker (1985) and a handful of subsequent studies have tried to gain some insight into the problem by considering a multiplicative error model where the observed extreme discharge data is assumed equal to the true extremes multiplied by a log-normally distributed random variable of unit expectancy and known variance. Hosking and Wallis (1997) adapted this approach in a regional setting. However, the multiplicative error-model approaches give an oversimplified description of true rating curve imprecision, which is essentially a curve fitting problem.

FRAMEWORK OF APPROACH

The estimation of a T-year return level using a regional approach consists of two steps: (1) the estimation of the rating curves used to calculate the flood discharges from measured flood stages, which in turns (2) are used to fit a chosen regional probability distribution. Consider a region Ω of R gauging stations. For any gauging station in Ω, say site j, there is a set of available stage-discharge measurements $\mathbf{D}_j = \left(Q_1^{(j)}, h_1^{(j)}\right), \ldots, \left(Q_{n^{(j)}}^{(j)}, h_{n^{(j)}}^{(j)}\right)$. Assuming a power-law rating curve model and multiplicative measurement error, one gets the regression model

$$q_i^{(j)} = a^{(j)} + b^{(j)} \log\left(h_i^{(j)} + c^{(j)}\right) + \varepsilon_i^{(j)} \qquad (1)$$

Where $q_i^{(j)} = \log\left(Q_i^{(j)}\right)$, i is an index going from 1 to the number of measurements, $n^{(j)}$, and $\theta_1^{(j)} = \left(a^{(j)}, b^{(j)}, c^{(j)}\right)$ are unknown parameters. Note that we here assume that the true stage-discharge relationship which covers the range of interest, i.e. all the flood stage values, is located in the upper rating curve segment. The measurement error $\varepsilon^{(j)}$ is assumed independent and normally distributed with zero expectancy and a finite standard deviation $\sigma^{(j)}$. The parameters $\theta_1^{(j)}$ which now include $\sigma^{(j)}$ can be estimated by maximising the associated likelihood function (Reitan and Petersen-Øverleir, 2006):

$$L\left(\theta^{(j)}\right) = \prod_{i=1}^{n^{(j)}} \frac{1}{\sqrt{2\pi}\sigma^{(j)}} \exp\left\{ \frac{\left[q_i^{(j)} - a^{(j)} - b^{(j)} \log\left(h_i^{(j)} + c^{(j)}\right)\right]^2}{2\left(\sigma^{(j)}\right)^2} \right\} \tag{2}$$

For a given stage S, the corresponding estimated discharge at station j is derived through the maximum likelihood (ML) estimated rating curve formula $\hat{Q}(S) = \exp\left(\hat{a}^{(j)}\right)\left(S + \hat{c}^{(j)}\right)^{\hat{b}^{(j)}}$. For example, the $m^{(j)}$ estimated annual maximum discharges at site j, $\mathbf{Z}^{(j)} = Z_1^{(j)}, \ldots, Z_{m^{(j)}}^{(j)}$, can then be calculated by applying the measured annual maximum stage values $\mathbf{S}^{(j)} = S_1^{(j)}, \ldots, S_{m^{(j)}}^{(j)}$ to the estimated power-law rating curve

Assume that the region Ω is statistically homogeneous with negligible intersite correlation, i.e. the R iid annual maximum discharge series follows the same distribution when normalised with the respective index floods $M^{(1)}, \ldots, M^{(R)}$. The expectation of the at-site un-normalised distribution, $E\left[Z^{(j)}\right]$, is typically used as the index flood. Furthermore, if it is assumed that this parent distribution is a GEV distribution, $F_{\theta_{\text{Reg}}}$, with regional parameters $\theta_{\text{Reg}} = \left(\xi_{\text{Reg}}, \beta_{\text{Reg}}, \mu_{\text{Reg}}\right)$ such that the at-site T-year flood is given by

$$x_T^{(j)} = M^{(j)} F_{\theta_{\mathbf{Reg}}}^{-1}\left(1 - 1/T\right) = M^{(j)} z_T \tag{3}$$

it is easy to see that the un-normalised annual maximum series at site j follows a GEV distribution with parameters $\theta_2^{(j)} = \left(\xi_{\mathbf{Reg}}, M^{(j)}\beta_{\mathbf{Reg}}, M^{(j)}\mu_{\mathbf{Reg}}\right)$.

The parameters θ_{Reg} can be estimated by the L-moments method. The expressions for the estimators of the L-moments and how to use these to estimate the GEV parameters are given in detail many places, e.g. Hosking and Wallis (1997). In short, one first computes the sample L-moment ratios for each site j in Ω using $\mathbf{Z}^{(j)} / M^{(j)}$, where $M^{(j)}$ is approximated by the sample mean of $\mathbf{Z}^{(j)}$. Then the pooled sample L-moment ratios are calculated as the average of these R ratios. Using the relationships between the theoretical L-moment ratios and the GEV parameters, and replacing the theoretical L-moment ratios with their sample counterparts, the system of equations can be readily solved for the GEV parameters.

Now that a framework for T-year flood estimation is available, an indication of the uncertainty associated with the estimate is needed. Invoking the delta-method, one obtains:

$$\mathrm{var}\left[\hat{x}_T^{(j)}\right] \approx \left(z_T\right)^2 \mathrm{var}\left(\hat{M}^{(j)}\right) + \left(M^{(j)}\right)^2 \mathrm{var}\left(\hat{z}_T\right) + 2M^{(j)} z_T \,\mathrm{cov}\left(M^{(j)}, z_T\right) \qquad (4)$$

This formula is difficult, if not impossible, to use in practice when both SV and RCI are considered. Instead one can apply the method of bootstrapping, which can routinely answer questions regarding the accuracy of an estimated statistic in situations far too complicated for traditional statistical techniques such as the delta-method. The basic bootstrap idea is quite simple: to emulate how the original data were constructed by replacing the true underlying distribution G with $\hat{G}$, where $\hat{G}$ is based on the observed sample. Re-samples of the data are drawn to build up a picture of the frequency of the statistic, then used for inference. Note that G might be a multivariate distribution. The bootstrap can be applied parametrically or non-parametrically. In the parametric setting $\hat{G}$ is represented by a parameterised distribution. The method also applies for problems where the data consist of several independent random variables, as in the present study. The theoretical background and technical details of the bootstrap method can be found in many textbooks (e.g. Efron and Tibshirani, 1983).

The procedure for parametrically bootstrapping the T-year flood at site j using the regional framework outlined in the former section is as follows:

Algorithm 1

For $u = 1,\ldots,U$

1. for $r = 1,\ldots,R$

   I. for $i = 1,\ldots,n^{(r)}$

      a. draw $e_i^{(r),*}$ from $N\left(0,\left(\hat{\sigma}^{(r)}\right)^2\right)$;

      b. set $h_i^{(r),*} = h_i^{(r)}$ and then set $q_i^{(r),*} = \hat{a}^{(r)} + \hat{b}^{(r)}\log\left(h_i^{(r),*} + \hat{c}^{(r)}\right) + e_i^{(r),*}$;

   II. use $\left(q_1^{(r),*}, h_1^{(r),*}\right)_u,\ldots,\left(q_{n^{(r)}}^{(r),*}, h_{n^{(r)}}^{(r),*}\right)_u$ to obtain ML estimates $\left(a^{(r),*}, b^{(r),*}, c^{(r),*}\right)$;

   III. draw $X_1^{(r),*},\ldots,X_{m^{(r)}}^{(r),*}$ from $\mathbf{GEV}\left(\xi_{\mathrm{Reg}}, \hat{M}^{(r)}\hat{\beta}_{\mathrm{Reg}}, \hat{M}^{(r)}\hat{\mu}_{\mathrm{Reg}}\right)$;

   IV. set $S_1^{(r),*},\ldots,S_{m^{(r)}}^{(r),*} = \left[X_1^{(r)}/\exp\left(\hat{a}^{(r)}\right)\right]^{1/\hat{b}^{(r)}} + \hat{c}^{(r)},\ldots,\left[X_{m^{(r)}}^{(r),*}/\exp\left(\hat{a}^{(r)}\right)\right]^{1/\hat{b}^{(r)}} + \hat{c}^{(r)}$;

   V. set $Z_1^{(r),*},\ldots,Z_{m^{(r)}}^{(r),*} = \exp\left(a^{(r),*}\right)\left(S_1^{(r),*} + c^{(r),*}\right)^{b^{(r),*}},\ldots,\exp\left(a^{(r),*}\right)\left(S_1^{(r),*} + c^{(r),*}\right)^{b^{(r),*}}$;

   VI. calculate $M^{(r),*}$ and $W_1^{(r),*},\ldots,W_{m^{(r)}}^{(r),*} = Z_1^{(r),*}/M^{(r),*},\ldots,Z_{m^{(r)}}^{(r),*}/M^{(r),*}$;

2. use $W_1^{(1),*},\ldots,W_{m^{(1)}}^{(1),*},\cdots,W_1^{(R),*},\ldots,W_{m^{(R)}}^{(R),*}$ to obtain L-moment estimate $\theta_{u,\mathrm{Reg}}^*$; then

3. calculate $x_{u,T}^{(j),*} = M^{(j),*}F_{\theta_{u,\mathrm{Reg}}^*}^{-1}\left(1 - 1/T\right)$

Dividing the sample standard deviation of the bootstrap sample $\mathbf{x}_T^{(j),*} = x_{1,T}^{(j),*},\ldots,x_{U,T}^{(j),*}$ with the corresponding sample mean, one obtains the coefficient of variation (CV) for the T-year flood estimate $x_T^{(j)}$, which is a fair measure of the estimation uncertainty, given that the bootstrap sample is reasonably symmetric. $\mathbf{x}_T^{(j),*}$ can also be used to approximate confidence limits. Testing of Algorithm 1 indicates that thirty thousand simulations are sufficient for achieving reasonable accuracy, in the sense that the Monte Carlo variability then is negligible compared to the various magnitudes approximated by the simulation. Note that by omitting some of the steps in Algorithm 1, one can exclusively study the impact of either SV or RCI.

APPLICATION

The instantaneous peak discharge will always be larger than the daily mean flood. When possible, one should therefore base a flood frequency analysis on instantaneous flood-peak values, use the digitally recorded, fine time-resolution, part of the time-series. However, since this typically commenced no earlier than the mid-eighties, such data series found on computerised databases are often short. A regional approach is then preferable.

Eight Norwegian gauging stations having virtually unregulated catchments were selected from the Norwegian hydrological database Hydra 2, held by the Norwegian Water Resources and Energy Directorate. Each time-series of annual maximum stage was carefully edited to remove years where the annual maximum was suspected to be missing of affected by backwater due to ice build-up. The rating curve segmentation limits were set in accordance with the official rating curves in Hydra 2.

The main characteristics of the stations are shown in Table 1. Standard regression theory suggests that several factors decide the degree of RCI: the degree of extrapolation, the number of stage-discharge measurements available and the magnitude of measurement noise, making it impossible to determine by a visual inspection prior to estimation. The degree of RCI at each site is therefore indicated by the CV for the median annual maximum discharge, $\mathbf{CV}(\hat{Q}_{\mathbf{MF}})$, calculated using a bootstrap scheme very similar to that given in Algorithm 1.

Hosking and Wallis (1997) present a method for determining whether a region of sites is homogeneous. It is based on a statistic H computed from the flood data. The behaviour of this statistic is compared with the "real" statistic obtained from simulations. If they have the same behaviour, the region is homogeneous. Hosking and Wallis (1997) suggest that the region be regarded as acceptably homogeneous if $H < 1$. For the present region Ω, H was 0.22 when RCI was neglected.

Table 1 presents the 50-year flood estimates from a regional analysis using all eight stations, while the CV for the 25- and 50-year floods for stations 027.015 and 027.016 are displayed in Fig. 1. One the x-axis, 1 is single-station estimation, while 2 is when these are pooled. Then, from 3 to 8, stations 027.025, 028.007, 027.026, 027.029, 027.020 and 026.032 are added, in this order. As expected, the uncertainty caused by SV is seen to quickly reduce as more sites are added to the region, before is stagnates. Even if it levels off at a value decided by the at-site index-flood variability (i.e. the first right-hand term in Eq. 4), a monotonic uncertainty reduction can be reasonably assumed, as more sites imply more data per parameter.

This is not the case for the uncertainty contribution from RCI. First, one sees from Table 1 that the uncertainty contribution from RCI can be large and significant, especially if the at-site RCI is large. Second, looking Fig. 1 the effect of RCI seems to change very little from its initial value as the number of stations increases, implying that the at-site RCI contribution contained in the index-flood dominates. Third, and most important, one sees that it exhibits both increasing and decreasing fluctuations around what appears to be an average value.

Table 1. Description of the eight gauging stations used in the analysis, and estimated 50-year floods with associated standard deviations (SD) where a, b and c represent uncertainty due to SV, RCI and the net effect of these two, respectively. The letters m and n represent the number of annual maximum instantaneous values and number of upper-segment stage-discharge measurements, respectively. CV(·) is the coefficient of variation of the median annual flood due to RCI.

Station	**Area (km^2)**	**m**	**n**	$\mathbf{CV}(\hat{Q}_{\mathbf{MF}})$	$\hat{x}_{50}$	$\mathbf{SD}(\hat{x}_{50})$
026.029	59.2	20	8	0.018	67.2	5.76[c] 5.01[a], 2.58[b]
026.032	79.0	20	8	0.025	74.0	5.95 5.58, 2.02
027.015	60.5	16	9	0.022	66.6	5.99 5.41, 2.30
027.016	124	22	12	0.065	232	24.1 17.1, 16.4
027.020	60.7	16	25	0.174	160	34.5 13.1, 30.5
027.025	645	19	12	0.009	468	38.0 36.1, 10.0
027.026	69.5	24	6	0.254	70.7	24.1 5.05, 21.5
028.007	140	20	8	0.055	100	10.2 7.62, 6.21

When a station with a very imprecise rating curve is added, i.e. 027.026, the total effect of RCI from all the stations exhibits an upwards trend. These observations have several implications. The net uncertainty of SV and RCI will be increasingly dictated by RCI as more stations are added to a regional analysis. At some point, RCI will surpass SV, as illustrated in Fig. 1c when the fifth station is included in Ω. In reality, this point will often be higher due to inter-site correlation. Another property introduced by RCI is the existence of a finite sub-region $\Theta \in \Omega$ which minimises the variability of a T-year flood estimate. Determining Θ analytically appears unfeasible, nor is subjective reasoning satisfactory. For example, if a preliminary analysis of a site *j* indicates a very high RCI, it might seem appropriate to exclude it. However, the degree of RCI and its impact in the frequency analysis are not easy to determine accurately prior to the flood frequency analysis. Even if they were, the assessment must also include the SV. For example, if the record for station j is particulary long, the total flood return level uncertainty might decrease by including it in the analysis, despite its large RCI. Numerical search techniques such as simulated annealing and genetic algorithms can theoretically be utilised for finding Θ, but if Ω contains a large number of stations, this will be extremely costly.

The characteristics of the bootstrap-samples $\mathbf{x}_T^{(j),*}$ merit some description. In most cases the bootstrap densities were symmetric, though a slight right-hand

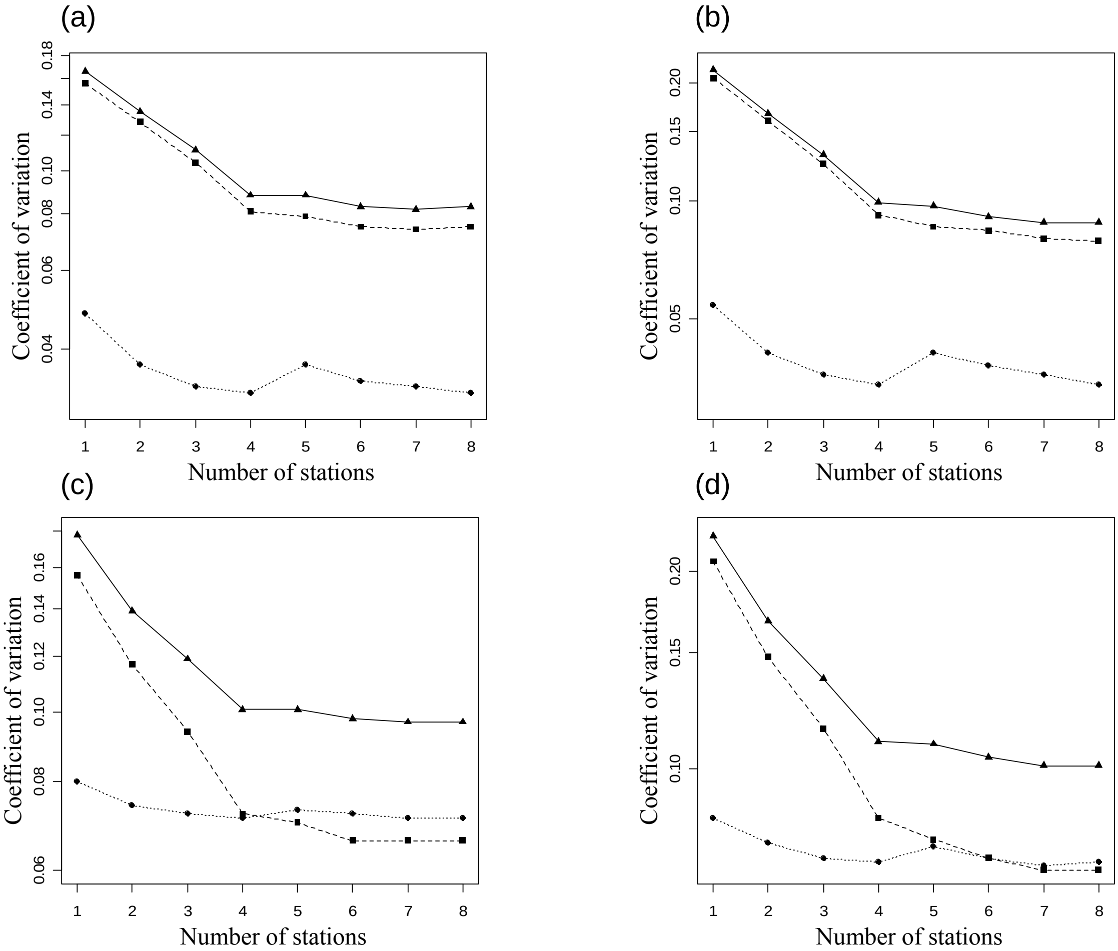

Figure 1. The coefficient of variation for (a) the 25-year flood, and (b) 50-year flood, for station 027.015, and (d) the 25-year flood, and (e) 50-year flood for station 027.016. Dotted lines with dots represent RCI, dashed lines with squares represent SA and solid lines with triangles are the net effect of RCI and SV.

skewness was observed in some samples. The sample mean of the various $\mathbf{x}_T^{(j),*}$ deviated little from the associated estimates. The largest deviation observed was less than 4 per cent. The symmetric and unbiased properties imply that Algorithm 1 can be used for accurate inference without invoking more advanced re-sampling strategies. It can also be inferred from Table 1 that the total variance can be well approximated by summing the variances of the two individual uncertainty sources. This is not very surprising, given that the normal assumption holds in both cases, since the SV and RCI are independent.

Finally, another interesting feature seen from Fig. 1 is that the impact of RCI seems to decrease, relatively to SV, with increasing return levels.

CONCLUSION

This study has developed a bootstrap-based framework for considering the coefficient of variation of a regionally-based T-year flood estimator based on the uncertainty contributions from sample variability and rating curve imprecision. Applying the methodology to data from a region comprising eight Norwegian gauging sites, it is shown that: (1) it is a computationally feasible procedure; (2) the inclusion of rating curve imprecision can lead to a significantly larger T-year flood estimation uncertainty; and (3) the uncertainty due to rating curve imprecision seems to immediately approach and then fluctuate around an average value as more sites are added to the regional analysis, in contrast to the effect of sample variability which decreases rapidly before stagnating as the station density grows. The latter implies the existence of a finite group, or sub-region, of sites within the region that minimises the T-year flood estimation uncertainty. However, the results must be qualified by the fact that they are only for one region of eight gauging stations and do not take into account inter-site correlation.

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ESTIMATING BAYESIAN RATING CURVES FOR THE ICELANDIC HYDROMETRIC NETWORK

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ABSTRACT

Hydrological rating curves are used to convert water level time series to discharge time series. The current method for the estimation of the rating curve is based on least squares. Its lack of in-depth uncertainty estimates causes problems in data analysis and it is virtually impossible to incorporate auxiliary information objectively into this method. The Chiefs of the Hydrological Institutes of the Nordic countries (CHIN) reported recently, Jónsson et al. (2002), that the subjectivity of the methodologies used by the member countries, causes uncertainties in the establishment of rating curves, resulting in different rating curve estimates between countries. Therefore, long term discharge averages and maximum values, based on the same data, are surprisingly different. This issue is addressed in a counter comparison with Bayesian rating curves in Árnason (2005).

Furthermore, an objective methodology for establishing hydrological rating curves based on Bayesian statistics is presented in general terms as well as the results of a thorough data research and analysis initiative, aimed at estimating Bayesian rating curves for the whole Icelandic hydrometric network and comparing them to the existing rating curves. Bayesian rating curves for 78 sites were constructed, using over 2800 discharge measurements, and compared to existing rating curves using Information Criterion methods as well as residual analysis.

The Bayesian approach naturally combines the statistical model for the data which incorporates the hydrological model, the data themselves and a priori information which is based on previously collected data and scientific knowledge. Data collected by the Hydrological Service in Iceland at the National Energy Authority are analyzed, using scientific and heuristic methods, establishing a priori knowledge about the required parameters. The combination of data and a priori knowledge results in a posterior distribution which is used to estimate parameters of the rating curve and to predict discharge for a given water level.

INTRODUCTION

The Icelandic hydrometric network can be largely divided into two groups depending on the location. The first group covers 34 stations located in the highlands (400 m or more above sea level). The other group includes the remaining 131 stations which are located under 400 m above sea level, Árnason (2005).

Around 85 of these stations are operated to provide discharge time series, which means that the water level is measured continously and converted to discharge using a hydrological rating curve. The rating curve is a statistical model, one of the most common being of the form:

$$Q = a(w-c)^b \tag{1}$$

Where $\boldsymbol{Q}$ is the discharge, $\boldsymbol{a}$ is a positive scaling parameter and $\boldsymbol{b}$ is a positive parameter containing information on the control section. The parameter $\boldsymbol{c}$ is the stage at zero discharge and is dependant on the local reference system of the hydrometric station. Defining the water level for a discharge measurement is an integral part of the measurement process. Usually, the weighted mean of the water level during the measurement is a sufficient estimate.

The main focus of this paper is to describe an objective methodology for establishing hydrological rating curves based on the Bayesian approach. Data collected by the Hydrological Service in Iceland (HS) is analyzed, using scientific and heuristic methods, establishing a priori knowledge about the required parameters. These data include all rating curve parameters from the network that have as a part of the project been gathered and organized into a database, functioning as the basis for Bayesian a priori information, as well as a plethora of other data, such as water level and discharge time series and the hydrometric network attributes. When producing discharge predictions, the Bayesian approach is utilized by using all these data from the HS.

The Bayesian approach naturally combines the statistical model for the data which incorporates the hydrological model, that data itself and the a priori information which is based on previously collected data and scientific knowledge. The result of this combination of information and knowledge results in a posterior distribution which is used to estimate parameters and to establish a rating curve as well as all derived data.

THE STATISTICAL MODEL FOR RATING CURVES

The classical statistical model usually used to describe the relationship between stage and discharge is given by

$$Q_i = a(w_i - c)^b + E(Q_i)\xi_i, \quad \xi_i \sim N(0,\sigma^2), \quad i = 1,\dots,n, \tag{2}$$

See Petersen-Øverleir (2004) og Moyeed and Clarke (2005), where (Q_i,w_i) is the i-th discharge measurement and stage for n measurements. Parameters $\boldsymbol{a}$, $\boldsymbol{b}$ og $\boldsymbol{c}$ are described following (1). The parameter $\boldsymbol{\sigma}^2$ is the variance of normal and independent residuals, $\boldsymbol{\xi}_i$, i = 1, …, n. The mean and variance of the discharge measurements in equation (2) is

$$E(Q_i) = a(w_i - c)^b, \quad Var(Q_i) = \sigma^2 E^2(Q_i) = \sigma^2 a^2 (w_i - c)^{2b}, \quad i = 1,\ldots,n.$$

The model given by (2) was improved upon by Petersen-Øverleir, see Petersen-Øverleir (2004), using a more flexible way to describe the variance, i.e.

$$Q_i = a(w_i - c)^b + E^{\psi}(Q_i)\xi_i, \quad \xi_i \sim N(0,\eta^2), \quad i = 1,\ldots,n, \tag{3}$$

Where the parameters $\boldsymbol{\eta}^2$ og $\boldsymbol{\psi}$ are used instead of $\boldsymbol{\sigma}^2$ to describe the variance as a function of $E(\boldsymbol{Q}_i)$. The mean and variance of discharge measurements in equation (3) are

$$E(Q_i) = a(w_i - c)^b, \quad Var(Q_i) = \eta^2 E^{2\psi}(Q_i) = \eta^2 a^{2\psi}(w_i - c)^{2b\psi}, \quad i = 1,\ldots,n.$$

The parameter $\boldsymbol{\psi}$ works thus; as $\boldsymbol{\psi}$ approaches 0, the variance of Q becomes fixed; if $\boldsymbol{\psi}$ = 1, the statistical model given by equation (2) is valid and the ratio of the standard deviation of $\boldsymbol{Q}$ and expected value of $\boldsymbol{Q}$, i.e.

$$CV = \sqrt{Var(Q)} / E(Q) = \eta a^{\psi-1}(w - c)^{b(\psi-1)} = \eta E^{(\psi-1)}(Q),$$

is fixed as a function of $E(\boldsymbol{Q})$; if $\boldsymbol{\psi} < 1$ CV decreases with increasing $E(\boldsymbol{Q})$; if $\boldsymbol{\psi} > 1$ CV increases with increasing $E(\boldsymbol{Q})$.

The parameters $\boldsymbol{a}$ og $\boldsymbol{\eta}^2$ in equation (3) are reparameterized, see Árnason (2005), with the parameters $\boldsymbol{\varepsilon}$ og $\boldsymbol{\tau}^2$, i.e.

$$a = \exp(\alpha_0 + \alpha_1 b + \varepsilon), \quad \eta^2 = \tau^2 a^{-2\psi}, \tag{4}$$

Where $\boldsymbol{\alpha}_0$ = 4,95 and $\boldsymbol{\alpha}_1$ = -5,37. This is done to diminish the effect of correlation in the posterior distribution of parameters $\boldsymbol{a}$ and $\boldsymbol{b}$. This reparameterization was first introduced in Árnason (2005). The variance and mean of discharge measurements in the reparameterized improved statistical model is

$$E(Q_i) = \exp(\alpha_0 + \alpha_1 b + \varepsilon)(w_i - c)^b, \quad Var(Q_i) = \tau^2 (w_i - c)^{2b\psi}, \quad i = 1,\ldots,n.$$

The statistical model given by equations (3) and (4) is, therefore, the model used to estimate the rating curves described and covered in this paper.

BAYESIAN STATISTICS

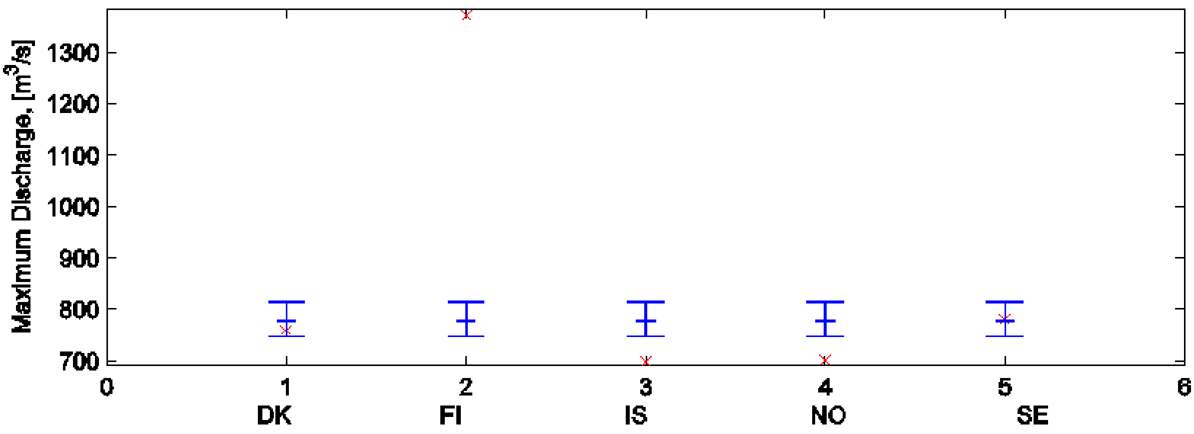
Bayesian statistics lay on the foundations of probability theory. All unknown parameters are treated as if they were random variables. The Bayesian approach requires a fully specified prior distribution for these unknown parameters, and a fully specified statistical model describing the observed data. The prior knowledge and the observed data are sources of information that the Bayesian approach naturally combines in a probabilistic framework. Its advantage is that all uncertainty can be taken into account, allowing for an accurate inference about the unknown parameters.

The use of Bayesian methods for fitting rating curves, is described in a recent paper by Moyeed and Clarke (2005). The Bayesian method is further improved upon in Árnason (2005) and that approach is used to derive the rating curves in this paper, a process which can be further referenced in Árnason (2005), Moyeed and Clarke (2005), Bernardo and Smith (1994) and Gelman et al. (2004). The estimation of posterior distributions includes, among other things, the use of a Gibbs sampler, the Metropolis–Hastings algorithm and the Markov Chain Monte Carlo (MCMC) sampling method. The model parameters are then given by the point estimate, i.e., the median values of each posterior distribution as well as credible regions for each parameter using percentiles of each distribution.

RATING CURVES AND CHIN

The Nordic countries have a formal agreement of cooperation under the acronym CHIN (Chiefs of The Hydrological Institutes in the Nordic countries).
The Nordic hydrological institutes have worked together on various issues related to their operations and research for a long time. This cooperation has, for example, taken form in workshops that are scheduled to investigate various issues concerning hydrological research. Leading experts from each institute are assigned to these work groups. In the late 1990's, a working group on rating curves was founded. That group published a paper on the uncertainties in the rating curve estimation process in August 2002, Jónsson et al. (2002).

Reenacting the analysis, a Bayesian rating curve was constructed for six rivers, using the same discharge data as the CHIN group. By using the innate strength of the Bayesian method, advanced credible regions, with similar properties as confidence intervals in classical statistics, were estimated for the annual runoff of the stations, for three different scenarios, and compared to the original findings. A sample of the comparison is shown in Fig. 1.



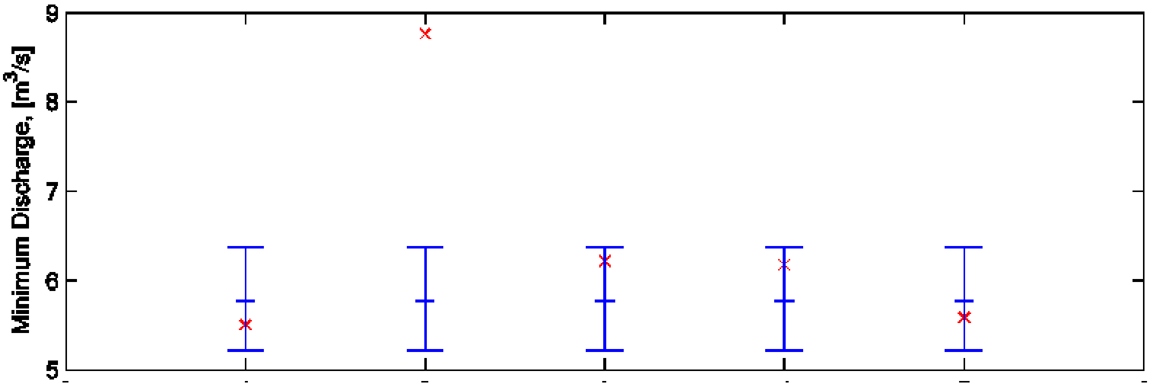


Figure 1. The estimate of maximum and minimum discharge for the Icelandic river Fnjóská for the sample year. The blue lines are Bayesian 95% credible regions with the estimated value of the maximum (upper figure) and minimum (lower figure) discharge in between. The red crosses are the maximum and minum estimates as reported by each CHIN member, Denmark, Finland, Iceland, Norway and Sweden.

In this case, the extreme discharge values calculated by the member nations for each station are compared to the Bayesian credible regions. It must be emphasized, that any notable difference in the value of the parameter ***b*** will result in vastly different discharge estimates if the rating curve needs to be extrapolated far. Furthermore, it is not entirely sufficient to compare the single segment Bayesian rating curve used here, to the multiple segment rating curves that are used in some cases for the CHIN results, as the multiple segment rating curves can have different ***b*** values for each segment. Given all these possible causes for difference, it is worth noting in Fig. 1 that almost all the member results fall close to or within the Bayesian credible regions.

COMPARING RATING CURVES

Discharge in an open channel is described by the aforementioned rating curve, based primarily on the channel shape, as dictated by the model parameters ***a***, ***b*** and ***c***. However, if the channel shape changes shape drastically with increased water level, a single rating curve may cease to describe the relationship of stage and discharge adequately. One accepted way to handle this event is to decide the water level where one rating curve is valid above and one below, each with its separate model parameters; $\boldsymbol{a_1}$, $\boldsymbol{a_2}$, $\boldsymbol{b_1}$, $\boldsymbol{b_2}$, $\boldsymbol{c_1}$ and $\boldsymbol{c_2}$, in which case the combined rating curve is called a double segmented rating curve with a breakpoint at water level $\boldsymbol{w_b}$. There is no limit to the number of segments, even though ideally, the segment number is usually constrained to a low number to reduce the chances of overfitting the data.

THE INFORMATION CRITERION METHOD

The methods premise is to compute a measure based on the sum of squared residuals (RSS), number of model parameters, $\boldsymbol{k}$, and number of data observations, $\boldsymbol{n}$. The measure rewards goodness of fit and includes a penalty that is a function of number of parameters. Therefore, the preferred model will be the one with the lowest IC value. There are a number of variations on the IC methodology with the Akaike's Information Criterion (AIC) and Bayesian Information Criterion (BIC) being the most widely used methods.

$$AIC = 2*k + n*\ln(RSS/n)$$
$$BIC = n*\ln(RSS/n)+k*\ln(n)$$

Samples of AIC and BIC calculations can be found in Table 1. The relative AIC and BIC values are both lower for the Bayesian rating curve compared to the existing rating curve, indicating that the Bayesian rating curve is the superior model of the two.

Table 1. Rating curve comparison using IC measures. A single segment rating curve has three parameters and a double segment rating curve has seven, six parameters and the breakpoint w_b. RSS is the residual sum of squares and the number of discharge measurements is 67. The rating curves are shown in Fig. 3 and Fig. 4.

IC parameters	*RSS*	*k*	*N*	*AIC*	*AICc*	*BIC*
IC values, Bayesian rating curve	28	3	67	-52	-51	-45
IC values, existing rating curve	22	7	67	-45	-43	-29

Additionally, robust residual analysis techniques and tests for residual normality are introduced in Árnason (2008), to further aid the objectivity of the estimation process.

THE BAYESIAN INITATIVE

All useful and unbiased data was compiled for every station in the Icelandic hydrometric network, including discharge measurements and water level extreme values. A single segment rating curve was then estimated for each station using the Bayesian approach described above. A sample result is shown in Fig. 2. and Fig. 3. whereas the existing rating curve is compared to the Bayesian rating curve in Fig. 4 and Fig.5.

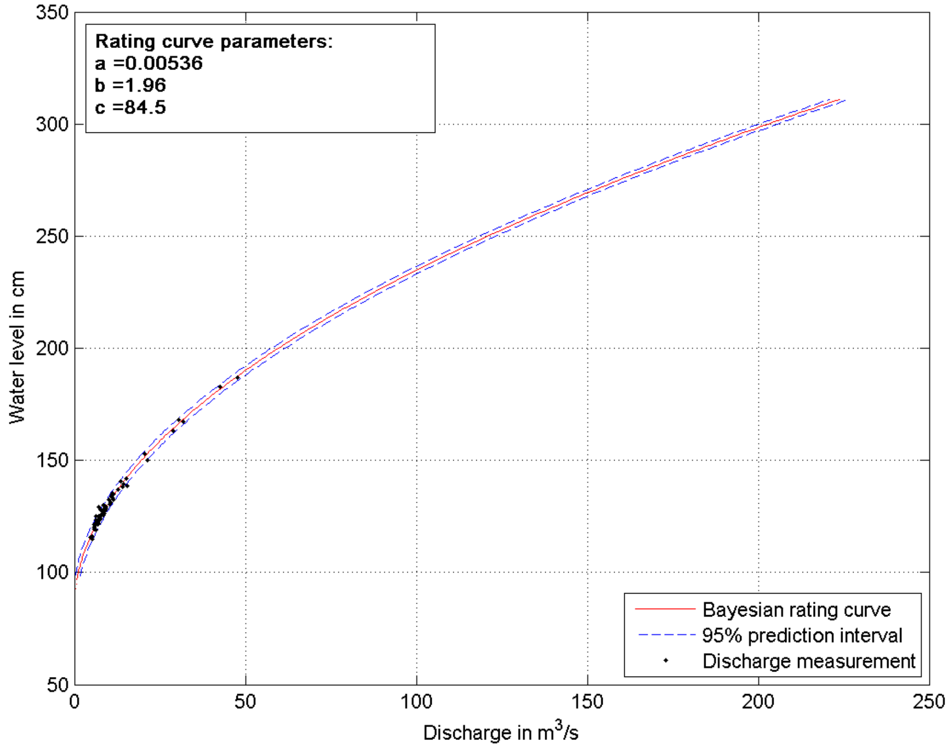


Figure 2. The Bayesian rating curve for hydrometric station VHM/V10 in the Icelandic river Svartá, Skagafirði. The range shown is from the extreme low and high values of the historical water level measured at the station. The broken line is the estimated 95% confidence interval for the discharge measurements.

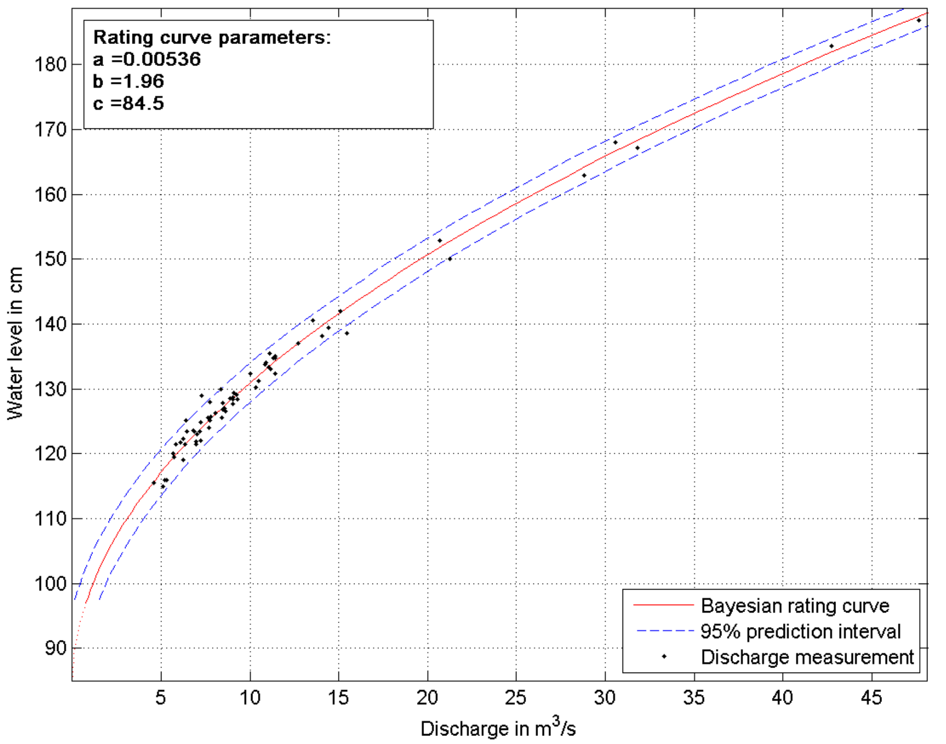


Figure 3. The same rating curves as in Fig. 2 with the water level shown contained to the highest and lowest discharge measurements.

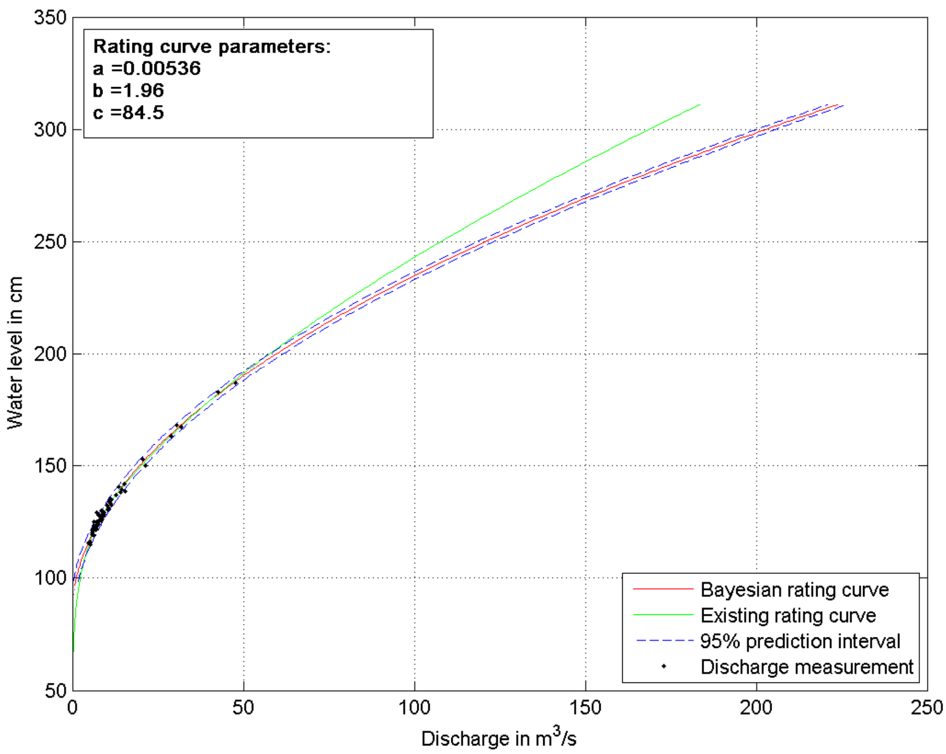


Figure 4. The Baysian rating curve and existing rating curve for Icelandic river Svartá, Skagafirði.

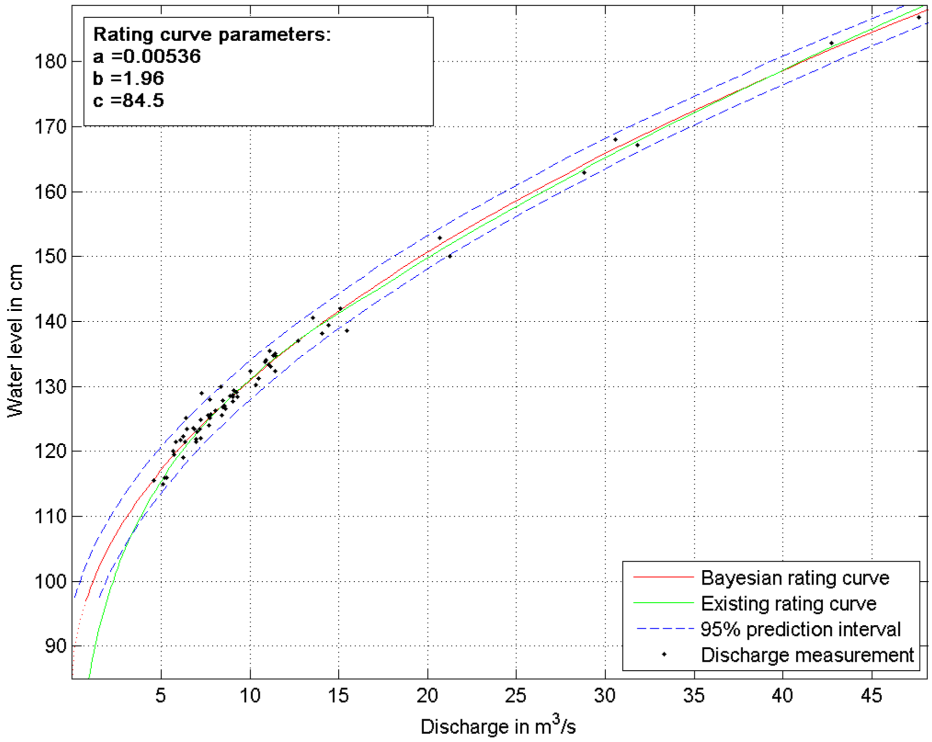


Figure 5. The same rating curves as in Fig. 4 with the water level shown contained to the highest and lowest discharge measurements.

RESULTS AND DISCUSSION

Prior studies have indicated that the current rating curves used at the Icelandic Hydrological Service may be subjective and suffer from overfit. To investigate this further and to remedy this problem, the HS launched a Bayesian initiative.

The Bayesian method described is both objective and unbiased and is believed to virtually eliminate this subjectivity. According to the results in Árnason (2008), the current Icelandic rating curves do indeed suffer from overfitting as measured by the Information Criterion method. Both the Akaike and Bayesian Information Criterion measures were calculated for every Bayesian rating curve and every existing rating curve and compared objectively as shown in Table 1. Of the 78 Bayesian rating curves that were calculated, 75 of them were deemed to be superior to the existing rating curves at the Hydrological Service.

There may be many explanations why this is the case but the fundamental difference seems to be that far to many segments have been used when estimating the current rating curves for the Icelandic hydrometric network. Three segments have been used when two segments might have been sufficient and two were used where one was sufficient. This theory can easily be verified when Bayesian double segment rating curves can be estimated for the whole hydrometric network and compared to the single segment rating curves using the same Information Criterion methods. Precisely that method is being described in a thesis currently in publication by Reitan and Petersen-Øverleir (2008) at the Norwegian Water Resources and Energy Directorate (NVE).

Judging by an estimate of residual scatter and the varied values of the variance parameter ψ, the new model introduced by Petersen-Øverleir (2004), deals efficiently with variance heteroscedasticity which is evident in the HS data and the results, establishing the functionality of the new model even further. In no case, does the proposed parameter of ψ equal 1, as it should, according to the traditional model. Furthermore, it is always less than 1, and usually differs by a large margin, clearly indicating a systematic error or bias in the traditional model. This is a critical fact and needs to be further addressed in future research as it indicates that the traditional statistical rating curve model may be limited and that the variance is described more successfully using Peterson-Överleirs new and improved model.

The results from the CHIN case studies are quite acceptable when compared to the Bayesian results. For almost all of the cases, the estimated annual means fall close to and symmetrically around the Bayesian 95% credible regions or within them. These annual mean values are as was mentioned before, calculated with different methods and either single or multiple segment rating curves. This indicates that the difference between the Nordic methods is not as drastic as was the initial concern. Nevertheless, these findings, including the results from Árnason (2008), give good reason to propose that the single segment Bayesian method is a valid method with the potential to become a common methodology for the Nordic countries. An advanced model, including shifts or multiple segment rating curves may, however, be necessary in exceptional cases.

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FLOOD RISK OF COASTAL AREAS TALLINN AND PÄRNU

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ABSTRACT

Over the last five years, Europe suffered over 100 major floods, including those along the Danube and Elbe rivers in summer, in the Alps in summer 2005, and along the Danube in the spring 2006 (EEA, 2007). Flooding of cities owing to spring flood and instantaneous rainstorms frequently occurs in Estonia as well. January 2005 the wind storm caused water rise in the Pärnu river and the inundation of the river mouth and the coastal area the city of Pärnu. The territory was suffered loss approximately 130 mln. US and 100 inhabitancies were affected (EM-DAT: The OFDA/CRED International Disaster Database, 2007). In order to improve flood protection in the EU Member States the Flood Directives should be implemented by 2015. Therefore, the rivers are discharged through the biggest cities Tallinn and Pärnu were selected as a pilot area for the study area. The preliminary flood risk assessment for these rivers has been done. Floods probability distribution 0.1% and 1% of water level was evaluated. The results of study will be used for the estimation of the flood risk and flood hazard maps, as the first step for implementation of the Flood Directives in Estonia.

INTRODUCTION

Flood is water rise, in the water level in a stream to a peak from which the water level recedes at a slower rate (WMO, 1992). Inundation or flooding is overflowing by water of the normal confines of a stream or other body of water or accumulation of water by drainage over areas, which are not normally submerged (Maastik, 2004). Over the last five years Europe was suffered over 1000 severe floods which brought huge damage to economy and have caused at least 700 deaths, the displacement of about half of million people and at least 25 billion Euro in insured economic losses (EEA, EMDAT, 2007).

The number of flood events were took place in Estonia as well during the last five years. There are different causes for the flood formation. Thus the coastal area of the biggest cities Pärnu, Haapsalu, Narva-Jõesuu and partly harbour area Muuga in Talinn were flooded in January 2001. The energy supply and phone lines were damaged by this storm. The South-West direction wind with speed - 20 m^3/s rise up the sea water level at the Pärnu city during a few hours by 1 meter. Thus, the absolute sea level record was exceeded in Tallinn (135 cm). However, this sea level was unofficial, because the data was taken from the automatic water level recorder, located at the harbour of Tallinn and maintained by the harbour. This sea level was not compared with homogeneous long-term time-series, measured by Estonian Meteorological and Hydrological Institute (EMHI).

The highest flood caused by strong storm damaged Pärnu, Haapsalu cities with surroundings, islands, and some North-West rivers (Keila, Vihterpalu, Kasari) of Estonia in 8-12 January 2005. The strong storm coincided with few factors - high amount precipitation, positive air temperature, water full storage of the soil and higher river discharge up to 130% and even for the Pärnu river basin up to 200% by the storm event. The probability distributions for these water levels are compared with 0.9-1.7%. The economical loss insured approximately 130 mln USD and 100 people were affected by storm in Pärnu (Kont, 2006).

Moreover, the absolute necessity for water level lowering of Lake Ülemiste appeared due to menace of inundation of Tallinn. The water level of lake was 37.03 mBS (Baltic System of Altitude), which did not reach the coastal fortification for 0.47 cm. Consequently, the increase of the water level in lake leads to the lost a capability for water intake from the discharged rivers Pirita, Keila, Jagala. Thus, the cellars of the surrounding houses were inundated.

Hydrological regime of rivers is typical for the East European Rivers with spring flood due to snowmelt and an autumn flood caused by rainfall and two low-flow periods. Local factors have a major influence to the river water regime such as land use – amount of lakes, bogs, forest; physical-geographical factors (slope), geological (tectonically breaks, paleohydrographical network, karst) and human impact – afflux from dams. The caused of inundation are different - spring flood due to intensive snowmelt and ice jam, flood caused by heavy rain or snowfall in autumn-winter and inundation at the coastal area, where the main factors are strong wind and high sea water level. Especially severe floods to evolved from overlapping of few such factors (Protasjeva, 1972).

During the highest spring flood 900 km^2 or 2% of Estonian area can be inundated. The water richest years were 1924, 1926 and 1931.

METHODS

There are few methods were used for the estimation of the probability distribution – Pearson type III and Kritsky-Menkel (Klibasev, 1970).

The Pearson type III frequency distribution was used for the study. This distribution was chosen because it's commonly used in Estonia and it has flexible coefficient of skew. In this method the annual maximum water level from the hydrometric station Pirita, for the period 1899-1995, and for the river Pärnu – Pärnu coastal station for the period 1923-2005 was used for the calculation.

The Pirita river discharges to the Gulf of Tallinn. The time series of the water level data from the Pirita river Kose coastal station was insufficient short – 1977-1987. Therefore, the data from the neighbouring coastal station Tallinn was used for the extension of the data. The correlation between the data set the Tallinn and Pirita stations was determined. Correlation coefficient 0.8 can be estimated as satisfactory and the data set for the period 1899-1995 was obtained.

The smooth empirical curve was constructed by this data, which allow determining probability distribution. Additional, for the confirmation of the obtained result the theoretical curve of frequency was constructed.

The theoretical curve should correspond to empirical curve and confirm the result. Thus, the computed higher water level with 0.1 and 1% frequency were found using binominal theoretical curve.

The Mann-Kendall test with probability 5% was used for investigation of the annual high water level trend (WMO, 1988).

STUDY AREA AND DATA

Two biggest cities – Tallinn and Pärnu, which are situated at the coastal of the Gulf of the Finland and the Gulf of Riga, were selected as a pilot area for the study. Advantage of the urban buildings and industrial constructions close to the seashore or close to the river is obvious. In this case capital investments for the hydro technical construction reduces, water head for enterprises and houses water supply are decrease, as well as the organisation for water-craft dock are reduced. Therefore, many flood-plains are build on and it is hazard of inundation, which related with human and material losses.

The water level regime at the study areas is determined by the influence of the sea and rivers. The greatest water level fluctuations determined by spring and autumn floods, rising tides and surge (fetch) – negative surge phenomenon. Surge phenomenon appear due to atmospheric pressure and wind to the water surface. At the study area the water level increases due to strong storm, which is related with an inundation of the coastal area.

The coastal station Pärnu is situated in two km from the river mouth Pärnu. The station is equipped by the tide gauge. The annual maximum water level data for the river Pärnu – Pärnu coastal station for the period 1923-2005 was used for the calculation.

Tallinn is a capital of Estonia. There are two different menace of inundation – coastal inundation due to the storm increase of the sea water level and potential menace of inundation from Lake Ülemiste, which is located in the city area (36.6 mBS). Protective barrage against the inundation is a half a meter higher, than the highest observed water level. Lake Ülemiste is the main source for the water supply of Tallinn. Except the water treatment plant there is also the emergency

Figure 1. The scheme of the study area – the Pirita river mouth in city Tallinn and the Pärnu river mouth in city Pärnu.

outlet from Lake. The catchment area extracts water from three rivers (Pirita, Jägala and Soodla) by means of a complex network of major reservoirs and channels covering approximately 4.5% of the Estonian territory.

At this study the potential inundation only for coastal area was estamated. Thus, the Pirita river, which discharges through the city of Tallinn to the Gulf of Tallinn was considered. The river mouth, which is the wide flood-plane with lengths 2 km is vulnerable for the surge phenomenon. However, at the flood plain the houses were build, communications and roads were constructed.

The annual maximum water level data from the hydrometric station Kose, which is located at 1 km from the river Pirita mouth for the period 1978-1987 was used for the calculation.

Therefore, that the water level data was insufficient short the data from the neighbouring station was used. The data from the coastal station Tallinn was used for the extension of the data. The distance between two stations is 4 km. The correlation between the data set the Tallinn and Kose stations was determined. Correlation coefficient 0.8 can be estimated as satisfactory and the data set for the period 1899-1995 was obtained. In 1996 the sea level measurements in Tallinn coastal station by the EMHI were discontinued. At this study only official data series from the EMHI were used for the calculations. It is quite possible, that during severe storms in recent decade sea level was higher, that recorded before, but the authors decided to take into account only homogeneous long-term series from EMHI. However, we have mentioned this data at the Table 1.

RESULTS

The empirical and theoretical curves were constructed for the maximum water level the Pirita and the Pärnu rivers. The result of the calculation is shown at the Table 1.

Table 1. The obtained coefficients variations and skewness for the maximum water level time series.

Coastal Station	***Period of observation***	***H** mean, **cm***	C_v	C_s
Kose	1899-1995	70	0.03	0.43
Pärnu	1923-2005	137	0.05	1.05

The Figure 2 shows theoretical and empirical frequency curves, which were obtained by the calculated parameters for the Pirita river. In this study it was found, that the constructed theoretical curve for *Cv=0.03* with *Cs=0.35* correspond to the observed data. This curve was used for the calculation. The theoretical curve of high water levels fit the empirical curve. They have an insignificant deviation from others observed water levels at the lower part of the curve. The highest water level 126 cm was observed in 1991, which is corresponding to 0.8 % probability. According to theoretical curve the water level - 142 cm correspond to 0.1% and 131 cm to 1.0% probability (Table 2).

The Figure 3 shows theoretical and empirical frequency curves, which were obtained by the calculated parameters. In this study it was found, that the constructed theoretical curve for *Cv=0.05* with *Cs=1.05* correspond to the observed data. This curve was used for the calculation. Two extreme high water levels lay above the curve for years 2005 and 1967. They have a significant deviation from others observed water levels. The extreme water level was 275 cm for year 2005 and 250 cm for year 1967.

The complex factors gave the result of the year 1967 flood; such as the deep Baltic Sea cyclone activity coincided with long sea wave. The Gulf of Pärnu is deeply sinks (22 km) inside to the main land. It has a narrow shape; therefore, in case of the wave set-up the outflow from the Gulf is difficult (Zahartsenko, 1988). Thus, during the flood 1967 the long wave filled the Gulf and additionally, the wave set-up added the water mass.

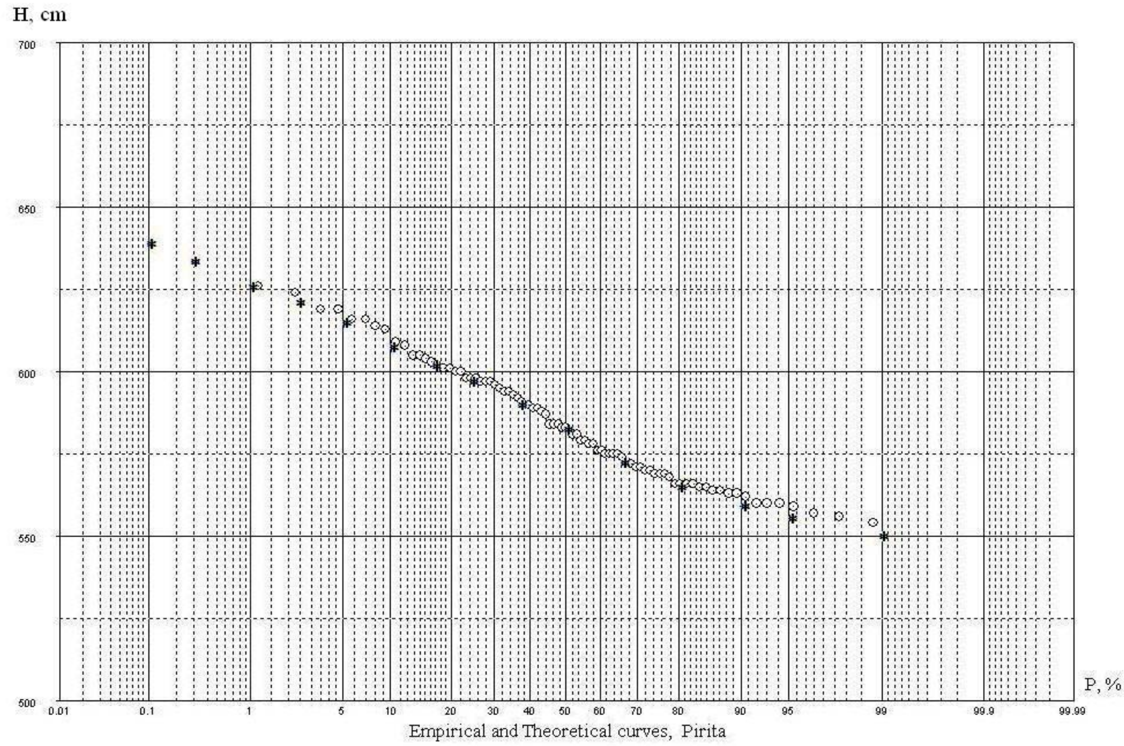

Figure 2. Empirical (filled dots) and theoretical (empty dots) curves of maximum water level of the Pirita river.

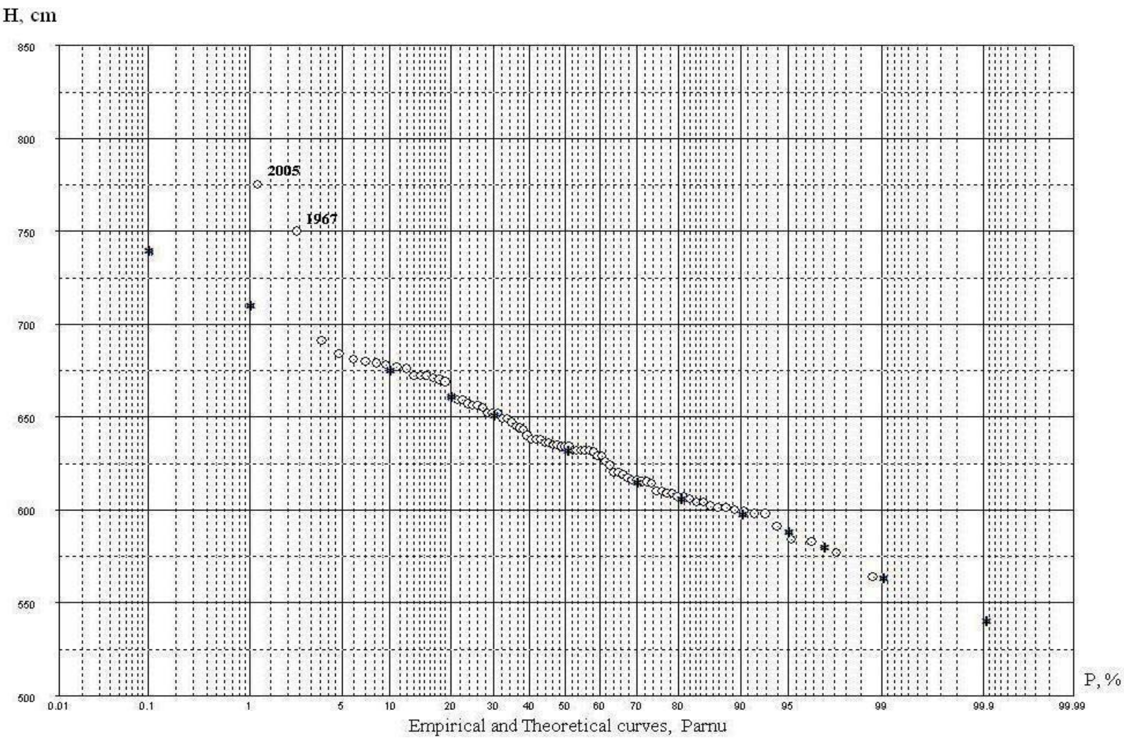

Figure 3. Empirical (filled dots) and theoretical (empty dots) curves of maximum water level of the Pärnu river.

Table 2. The obtained maximum water level with 0.1% and 1.0% probability.

Coastal Station	Period of observation	H_{max}, cm by EMHI	H_{max}, cm unofficial	Probability 0.1%	distribution 1.0%
Kose	1899-1995	110	149 (2005)	142	131
Pärnu	1923-2005	275		239	213

The next observed level correspond to 191 cm in year 1969. The main cause for the highest historical water levels in 2005 and 1967 was rising of the water level by the effect of South West wind with maximum speed by 34 m/s and up to 48 m/s. The high water level was observed at all coastal stations of the Gulf of Riga.

If take into account the cyclical variation of the highest floods, it can be assumed, that the frequency of the highest floods as years 1967 and 2005 is less, than was determined by empirical equation. Thus, if move the position of the highest water levels 2005 and 1967 to the empirical curve with *Cv=0.05*, the probability of the highest water level come to 0.01% and 0.04%. However, according to empirical equation the probability of these levels is compared with 0.8% and 2.0%.

The Figure 4 shows the long-term time series of the annual higher water level of the Rivers Pärnu and Pirita. The magnitude of fluctuations of the water level is higher for the Pärnu river. The Pärnu river belongs to the rivers, discharged into

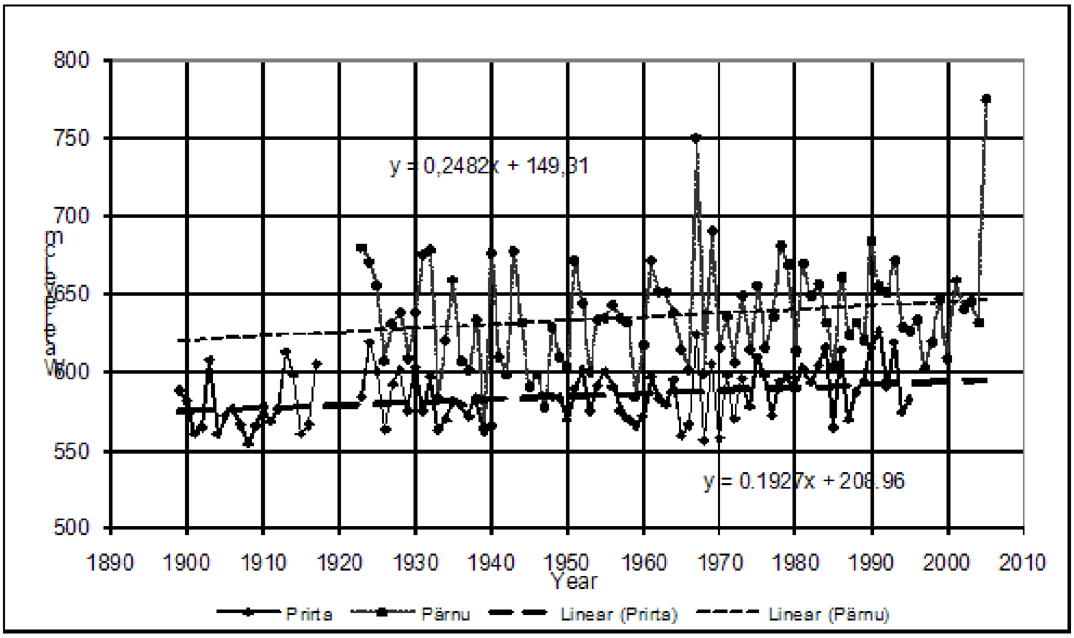

Figure 4. Long-term annual maximum water level with trends for the Pärnu and Pirita rivers.

the Gulf of Riga. According to investigations (Kalje, 1970) the river has strong influence to the water level of the coastal part of the Gulf of Riga, therefore, the magnitude of the fluctuations water level of the Pärnu coastal station is higher, than at the coastal part of Tallinn (the Pirita river).

The linear trends of the maximum water levels are increasing for both stations. The higher trend was obtained for the Pärnu river, where the annual water level rise was 2.2 mm, at the same time, the slightly lower increase – 1.9 mm was noted for the Pirita river. However, the increasing linear regression trend is significant at the level 1 % for the Pirita river. Moreover, the increasing trend was confirmed by the Mann-Kendall test at the significant level 5 %. However, no any significant trend was found for the Pärnu river.

This result confirms the results, which were obtained by Estonian scientists-geologists (Petersell, 2008) that the coastal part of Estonia and particulary the Gulf of Finland is under the slight tectonically down-lift. However, the magnitude of the coastal lifting can, probably, mean, that there is the eustatic increasing of the water level, but not the tectonical vertical movement of the coastal area.

Thus, increasing trend is mostly related to the global warming and consequently, with ice melting and the increasing of the sea water level.

CONCLUSION

Floods probability distribution 0.1% and 1% of water level was evaluated for the Pirita and Pärnu rivers. The results of study will be used for charting of the flood risk and flood hazard maps, as the first step for implementation of the Flood Directives in Estonia, which allow to determine flooding area and consequently, to forecast and assess possible losses and damage.

The linear trend of the maximum sea level with significant 1% shows the increasing trend, which means the increasing of flood risk for the city of Tallinn. Even, that the trend is not significant at the Pärnu station it has also growing tendency.

Moreover, the study shows, that the renewal of the measurements at the Pirita river mouth is desirable for the flood forecast and control, due to a high risk of damage for the coastal area of Tallinn.

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EXTREME RAINFALLS AND DAMAGES ON AUGUST 13 2007 IN THE CITY OF TRONDHEIM, NORWAY

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ABSTRACT

This paper is on the study of a frontal rainfall and flash flood that occurred in the City of Trondheim August 13 2007, causing one of the worst flooding in the City. More than 100 basements were flooded resulting in tremendous damages. Data from six short term rainfall gauges were analyzed. They were compared with the data from the weather radar at Olsøyheia in Rissa 30-40 km north of Trondheim center. There was a good correlation between the data from the rainfall gauges and from the radar. The front of the rainfall hit Saupstad at 18:15 and Risvollan and Voll almost at the same time at 18.22. Than it hit Ranheim at 18.29 and Lade at 18:31. Sverresborg at the west side of the city was hit at 18:26, with the highest intensity at Risvollan with 550l/sha and lowest at Ranheim with 206l/sha. The radar shows that the rain front moved about 8 km during 5 minutes (32 km/t = 9 m/s). The most intensive part of the rainfall had duration of 5 minutes and had a spatial distribution on 2.5 – 3.0 km. The biggest total rain depth is in the south, while the lowest total depth is in the northeast, the Ranheim area.

The project group has concluded, that to avoid flooding, is it important to safeguard surface floodways for the extreme rainwater/stormwater runoff and to make the urban drainage systems more efficient.

Keywords: Extreme rainfall, rainfall analyzes, rainfall intensity, rainfall return periods, rainfall movement, short-term rainfall.

INTRODUCTION

This study is on the frontal rain and the flash flood that occurred in the City of Trondheim on August 13, 2007, between 07:00 and 08:00 PM, local time, (18:00 – 19:00 normal time), causing one of the worst floods in the City of Trondheim (Fig 1).

The rainfall front moved in from the Atlantic Ocean, hitting first the southern western part of the city in the Heimdal region. Then it moved from southwest to northeast causing huge damages all over the city, i.e. floods in more than 100 basements (Fig. 1). Floods in basements represent the movement of the rainfall over the City.

The urban hydrological measuring and research station at Risvollan, located at 63° 24' N and 10° 25' E, at 85 masl, 4km south-east of downtown Trondheim, see, www.ivt.ntnu.no/ivm/risvollan/, was also flooded, so runoff data were lost, Thorolfsson (2007)

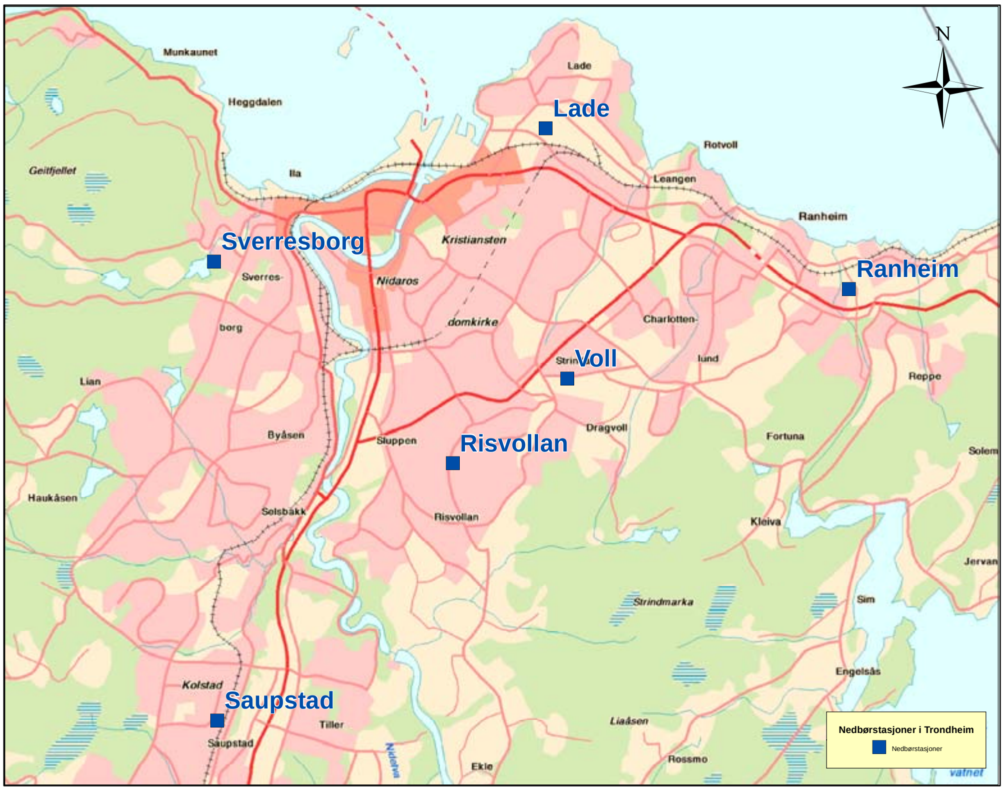

Figure 1. Locations of the six short-term rainfall gauges in the City of Trondheim.

The City of Trondheim has established a project group whose task is to for analyze this extreme rain event and to clarify the causes behind the tremendous damages in the City. The authors of this paper are the members of this group.

RAINFALL MEASUREMENTS

The short-term precipitation is measured in Trondheim at six locations, i.e. Saupstad, Sverresborg, Risvollan, Voll, Lade and Ranheim (Fig. 1, so good registrations on the rain intensity are available on August 13, 2008. The highest rain intensity at Risvollan was measured to 3.3mm/minute, equal to 550l/sha, the highest ever measured in Trondheim.

The short term intensity has been recorded in Trondheim since 1905. From 1967 the gauge was located at Tyholt, than from 1988 located at Moholt and finally located at Voll. Totally there are 36 measuring seasons available. The IDF-curves based on these measurements are used today for dimensioning stormwater facilities in the City of Trondheim (Fig. 2), while Risvollan has been collecting data since 1986, i.e. over 20 seasons.

ANALYSING RAINFALLS

Table 1 shows the intensity for 1, 5, 10, 15 and 20 minutes rainfalls at the six rainfall gauges.

Table 1. The intensity for 1, 5, 10, 15 and 20 minutes rains at the six rainfall gauges

Location	**1min** l/s*ha	**Voll** year	**5min** l/s*ha	**Voll** year	**10min** l/s*ha	**Voll** year	**15min** l/s*ha	**Voll** year	**20min** l/s*ha	**Voll** year
Ranheim	206	4	151	9	84	3	57	2	57	2
Lade	273	15	215	76	128	29	89	14	89	14
Voll	367	100	250	>100	147	77	100	27	77	18
Risvollan	550	>100	327	>100	200	>100	139	>100	139	>100
Sverresborg	300	24	260	>100	168	>100	118	100	118	100
Saupstad	300	24	240	>100	185	>100	141	>100	141	>100

The table shows great variations in the rain intensity according to the location. At Risvollan, Sverresborg and Saupstad the 1, 5, 10 20 and 20 minutes intensities, have return period over 100 years for all durations, while Voll, 3 km northeast of Risvollan has return periods on >100, 77 and 27 years. Ranheim in the northern part of the city has the lowest return periods with only 9, 3 and 2 years.

The rain intensities for 1, 5, 10 and 15 minutes are plotted into the IDF – curves for Tyholt – Moholt – Voll (Fig. 2).

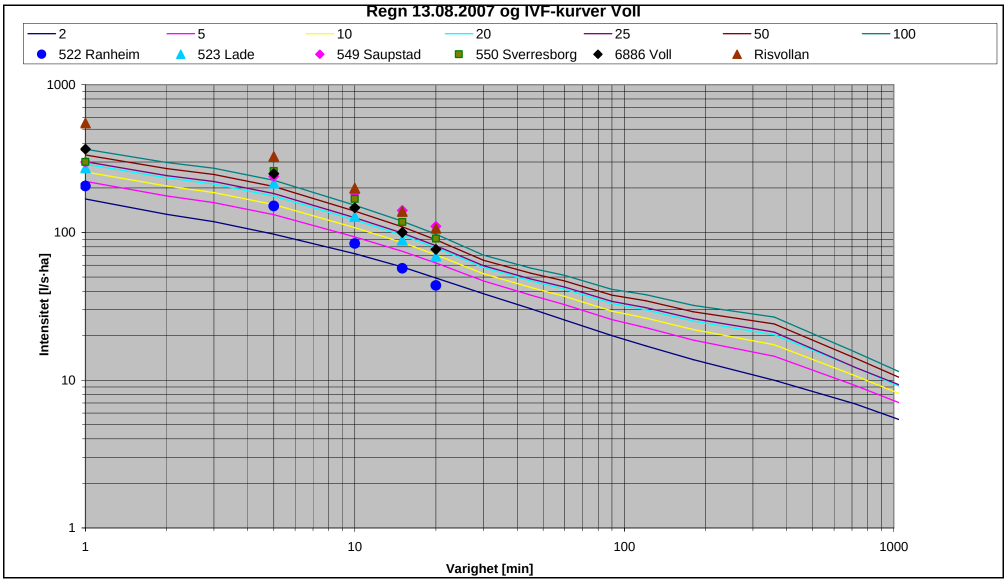

Figure 2. IDF – curves for Voll with corresponding rain intensities and return periods for Ranheim, Lade, Saupstad, Sverresborg, Voll and Risvollan on August 13 2007.

According to the recommendation on designing stormwater facilities, the return period should be at least 10 years for residential areas and 20 years for city centers, commercial areas and industry areas, Lindholm et al. (2005). Earlier practice was to use 10 or 5 years return period, and even down to 2 years, Lindholm et al. (2005).

The weather radar located at Olsøyheia in the municipality of Rissa, located ca. 30-40 km north of Trondheim, has recorded the movement of the rain front on August 13, 2008 over the City of Trondheim, showing the movement of the rainfall (Fig. 3). Figure 3 shows that the rainfall moved from southwest at ca. 18:00 PM and than moved over the central part of the City of Trondheim. It had passed at 19:00 PM, one hour later. This is in accordance with the measurements at the six rainfall gauges.

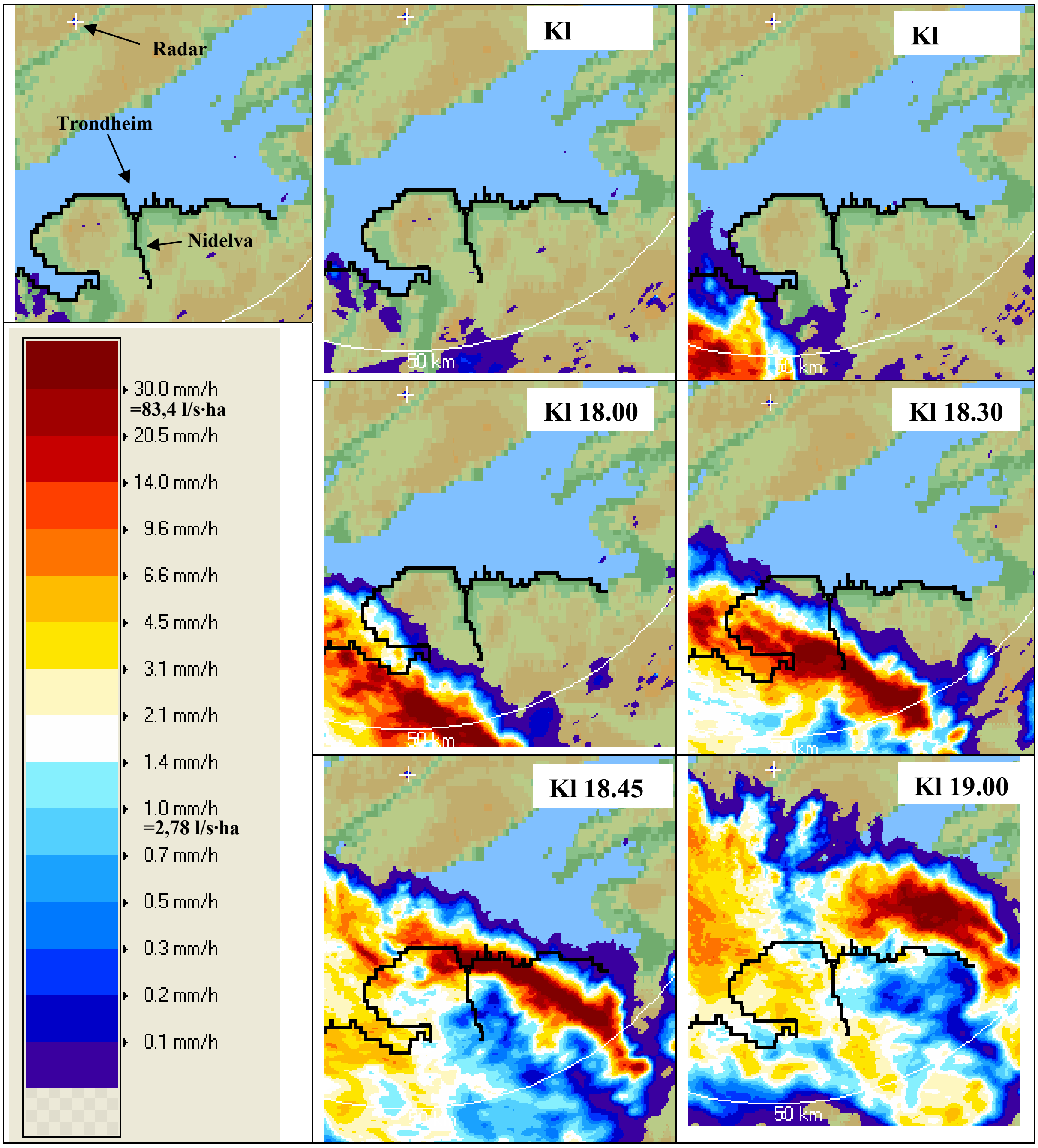

Figure 3. The movement of rain and rain intensity over Trondheim on August 13, 2007 at 18:00 – 19.00 normal time i.e. 07:00 – 08:00 LT. (Photos from *www.met.no)* *http://retro.met.no/radar/trondelag.html*.

Fig. 4 shows measured rain intensity at all six measuring stations between 18:00 and 19:00 PM.

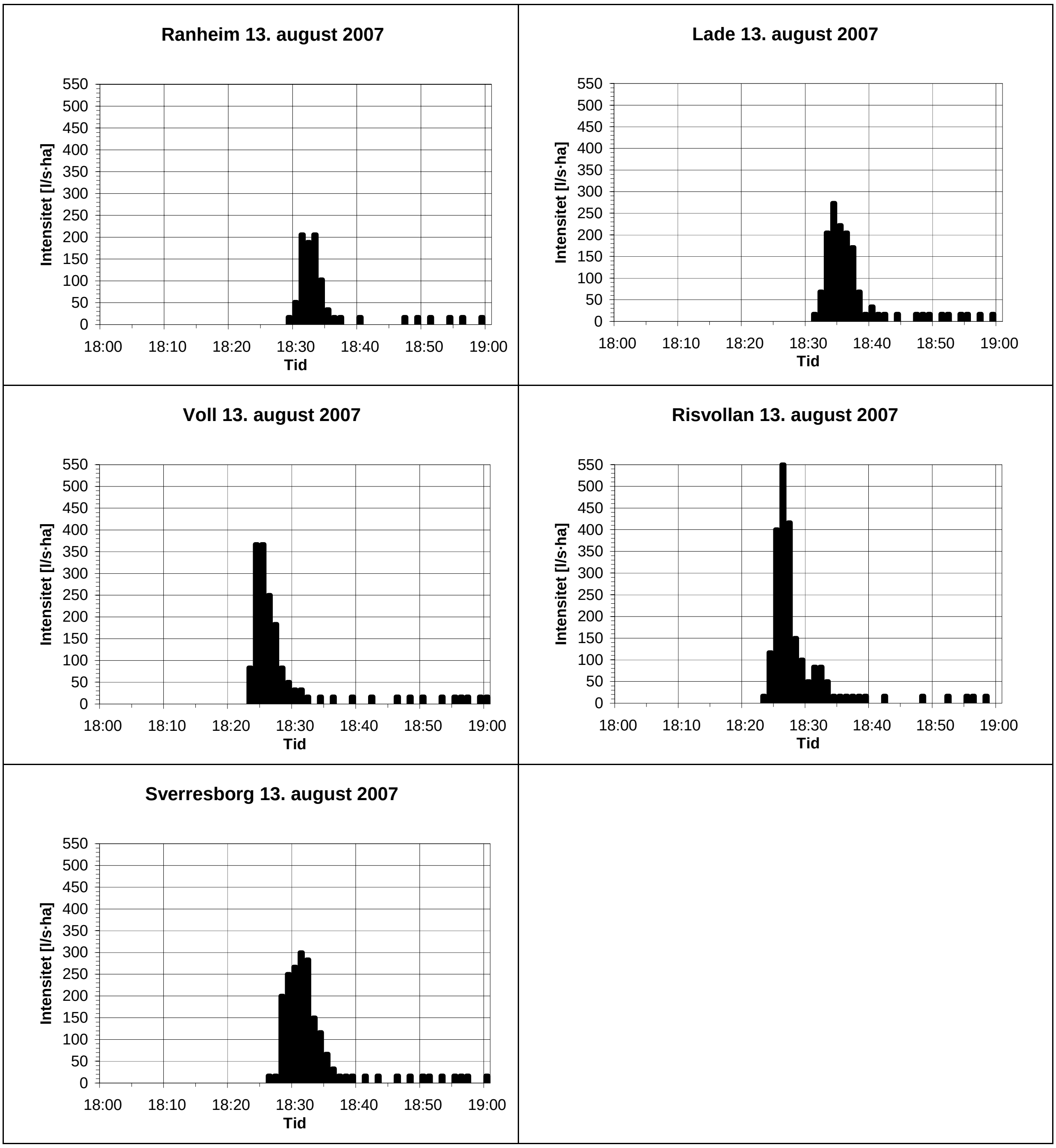

Figure 4. Measured rain intensity in the City of Trondheim August 13 2007 (time in normal time), Risholt (2007)

DISCUSSION

The front of the rainfall is hitting Saupstad at 18:15, than hitting Risvollan and Voll almost at the same time 18.22, than hitting Ranheim at 18.29 and Lade at 18:31. Sverresborg at the west side of the city was hit at 18:26, with the highest intensity at Risvollan with 550l/sha and lowest at Ranheim with 206l/sha.

The pictures from the radar show how the rain front moved about 8 km during 5 minutes (32 km/t = 9 m/s). The most intensive part of the rainfall had duration of 5 minutes and had a spatial distribution on 2.5 – 3.0 km.

The biggest total rain depth is in the south, while the lowest total depth is in the northeast, the Ranheim area. Only half of that amount was falling.

The municipality of Trondheim has established a project group for analyzing this extreme rain event to clarify the causes behind the tremendous damages in the City

According to recommendation on dimensioning of stormwater facilities, the return period should be at least 20 years, Lindholm et al. (2005).

CONCLUSION

On August 13 2008 at 18:00 – 19:00 PM, a heavy rainfall moved from southwest to northeast over the City of Trondheim with 14.6 mm as the biggest value.

This rainfall caused great damages all over the city, but mostly in the path of the rainfall that was located between the six rain gauges at Saupstad in south, Sverresborg in west, Risvollan and Voll in east and Ranheim and Lade in north.

A rainfall with duration on 5 minutes is giving the longest return period, i.e. over 100 years for Saupstad, Sverresborg and Risvollan, 76 years for Lade and nine years for Ranheim. All damages have occurred in areas with rain intensities with over 60 years return period, i.e. considerably higher than the dimensioning values, Lindholm et al. (2005).

The project group has concluded that to avoid flooding, is it important to safeguard surface floodways for the extreme rainwater/stormwater runoff and to make the urban drainage systems more efficient.

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ON THE ROAD TO A NEW NATIONAL HYDROLOGICAL DATABASE OF ICELAND

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ABSTRACT

The Hydrological Service in Iceland is working on a set of digital spatial data that contains information about surface water features. The objective will be a nationwide coverage of a direction-based hydrological data in the scale of ca. 1:50,000 with greater accuracy and detail than existing database. The database will be used for future work on the adaptation of the European Water Framework Directive. This paper lists the origin of the data provided and describes problems with processing a very high-resolution hydrographical database.

INTRODUCTION

In 2007, the Icelandic parliment voted for adaptation of the Water Framework Directive (WFD) of EU (http://www.euwfd.com) with the objective to fulfill its requirements before 2017. The initial work to comply with its adaptation is to define watersheds and the so-called water bodies, which will subsequently be characterized for pressures, impacts and economic analysis.

The existing hydrology database in Iceland is comprised of two parts:1) a digital image of rivers and lakes in a vectorised and georeferenced ArcInfo coverage in 1:250,000 scale from the Iceland Geodetic Survey (IGS) with classification of lines and polygons according to standards at IGS and the National Energy Authority/Hydrological Service (NEA/HS); and 2) an Oracle database edited at the HS, which includes separate tables of river and lake id‘s as well as other attributes, such as inter-relationships between rivers/lakes, (Jóhannsson and Einarsson, 1998). This combined database is a spaghetti vector data model with no flow directions, or river center-lines.

A study was made on a subset of the 1:250,000 hydrology database to estimate the time and cost needed to adjust it to the 1:50,000 IS50V water database of IGS and to add data such as flow directions and center-lines required for further work for the WFD adaptation. The main results of this study were that the cost for this modification exceeded available funds and that the IS50V water database was inadequate in many instances.

Hence, to continue work for the WFD adaptation, the NEA has leased a high-resolution (<1:5000) digital database of surface water features from Loftmyndir ehf., including lakes, ponds, streams, and rivers. This database includes attributes such as center-lines and flow directions required for watershed establishment and future work for the WFD. For the last year, the HS has worked on this digital database in order to correct and modify it for future use. The final product will be a nationwide coverage of hydrological data in the scale of 1:50,000; replacing the 1:250,000 data base. The advances to the available database will be a topologically correct dataset and tabular data storing directions, names (river and lake id‘s) and connections to water bodies making it possible to build a river network in the future. A river network database enables one to trace water movement through the landscape.

ORIGIN AND STRUCTURE OF THE HYDROLOGICAL DATABASE

The database from Loftmyndir ehf. is comprised of digitized polygons and lines representing various hydrographic features such as rivers, lakes, shorelines etc. shown in Table 1. The lines were traced from aerial photographs covering the entire country with up to 1 m accuracy. A Digital Elevation Model (DEM) is needed before watersheds can be delineated therefore, a DEM with a 25-m-pixel size and 10-50 m accuracy was obtained from Iceland GeoSurvey (ÍSOR) in 2007. The DEM was constructed combining from different sources the most accurate elevation data available. As a result the quality of the DEM varies with location for instance the area around glaciers and their vicinity has the poorest quality due to lack of data.

The structure of the national hydrological database will be based on the ArcHydro data model schema (Maidment, 2002), storing the existing 1:250,000 hydrological river-id from the older database (Jóhannsson and Einarsson, 1998). The ArcHydro data model was chosen because it has become one of the most tried and tested GIS data models for water resources. It provides a simple way of linking time series data to geospatial data within a single information system making it possible to trace water movement throughout the stream and river network.

Table 1. Main polygons and lines in the Loftmyndir database.

Polygons	**Lines**
Glaciers	Glaciers
Hot springs area, fumaroles	Hot springs
Intermitted lakes	Hot springs area, fumaroles
Foreshore flat	Intermitted lakes
Bare rocks and islands in rivers and lakes	Coastline (low tide)
Bare rocks and islands in sea	Bare rocks and islands in rivers and lakes
Lake	Bare rocks and islands in sea
River (<7 m)	Coastline (high tide)
Mud area and intermitted creek area	Wharf
	Pier
	Ditches
	Intermitted rivers
	Lake
	Riverbank
	River, centerline
	Mud area (unvegetated)/intermitted creek area
	Rapids
	Waterfalls
	River, centerline rivers <7 m
	Connecting line btw. rivers/ditches etc. and centerlines
	Dams
	Culvert

WORK PHASES AND PROCEDURES

The objective is to create a layer consisting of direction-based hydrological data in the scale of 1:50,000. In order to do this the data has to be processed to meet the requirements of the project. Figure 1 pictures the main phases of the project.

Data preparation

Various errors were encountered in the initial error-check of the digital data set from Loftmyndir and the DEM from ÍSOR.

1) The traced lines used in the Loftmyndir dataset were produced using a CAD Microstation system which has produced problems in converting the data to ArcGIS. Therefore, time-consuming correction of topology has been carried out in the pre-processing phase. Basic topology rules were defined so that lines were for example not allowed to self overlap or intersect (Figure 1).

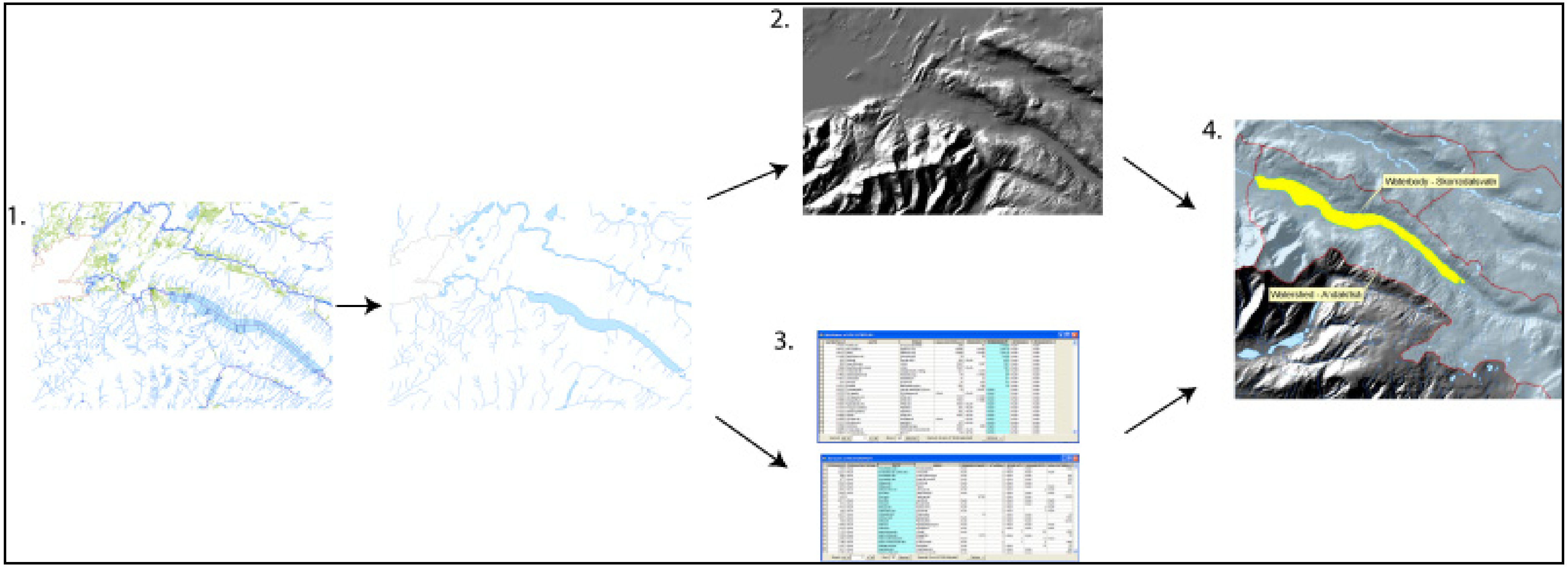

Figure 1. The main phases of the project: 1. Generalization of the Loftmyndir database. 2. DEM perperation. 3. Attribute transfer. 4. New hydrological database with waterbodies for WDF.

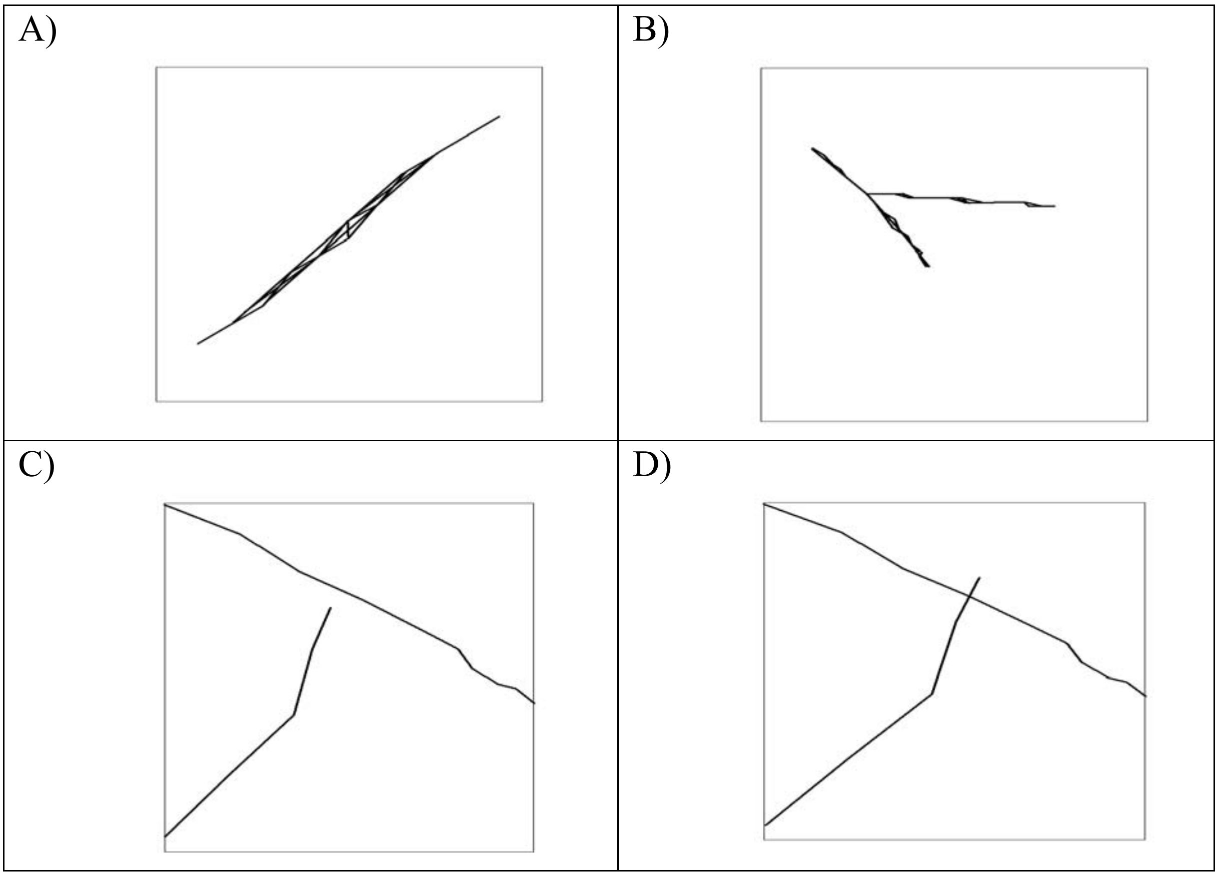

Figure 2. Different types of errors found in the Loftmyndir data set which have to be correcting within the pre-processing phase. A) Intersect; B) Overlap; C) Undershoot; and D) Overshoot.

2) The resolution of the traced lines is too high for subsequent work in the dataset, e.g. at present every line in the dataset, such as a ditch or a river, has equal importance. Thus, a simplification is needed before it is possible to work with the data set at the intended scale of 1:50,000. A Strahler classification will be attempted on the database and subsequently used to simplify it. For the Strahler classification the RivEx tool (see www.rivex.co.uk), an add on program that works with ArcGIS, will be used.
3) The DEM is only approximate representation of land surface terrain. In order to create an accurate representation of flow direction a depressionless DEM is needed. Sinkholes need to be filled and streams need to be burned in to allow for correct flow direction calculations.

Delineation of watersheds

Combining the classified/simplified data set from Loftmyndir and the DEM data set has been tested for individual areas with promising results. For the results to be meaningful careful preparation of both the hydrographic database and the DEM needs to be completed.

Connection to the older database of rivers and lakes

Moving the old river and lake id‘s from the existing 1:250,000 database has to be done manually due to the much higher resolution of the Loftmyndir dataset and inaccuracy in the older database. The old database consists of more than 5000 rivers-id’s and more than 2000 lakes with names that have to be transferred (Jóhannsson and Einarsson, 1998).

CONCLUDING REMARKS

The final deliverables of this work will be a national direction-based hydrological database in the scale of ca. 1:50,000, which will be the basis for watershed definition and other prerequisites for the adaptation of the European Water Framework Directive. The hydrological database will have a national coverage and a greater accuracy and detail than the database of rivers and lakes already existing at the Hydrological Service. This work is still ongoing with the initial phase being finished in 2008.

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BAYESIAN ESTIMATION OF DISCHARGE RATING CURVES

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ABSTRACT

The Bayesian approach has been successfully applied to the estimation of discharge rating curves which are based on the standard power-law. Here the standard power-law model is extended by adding a B-spline function to it. The extended model is compared to the standard power-law model through discharge data from the direct run stream Norðurá in Borgarfjörður in the Western part of Iceland. The extended model provides a substantially better fit to these data than the standard power-law model.

INTRODUCTION

Hydrological Service in Iceland (HSI) runs a water level measuring system which collects water level data continuously from rivers around the country while the discharge is only measured a few times a year due to high cost. Hydrological rating curves give discharge as a function of water level. Based on hydraulic principles, the relationship between discharge and water level is given by the standard power-law relationship

$$q = a(w - c)^b \tag{1}$$

(Lambie, 1978; Mosley & McKerchar, 1993) where q is discharge, w is water level, a is a positive scaling parameter, b is a positive shape parameter and c is the water level when the discharge is zero. These parameters are usually estimated from paired measurements of water level and discharge.

The Bayesian approach has been successfully applied to discharge rating curves, see Moyeed & Clarke (2005), Reitan & Petersen-Øverleir (2008) and Arnason (2005). In the Bayesian approach all unknown parameters are treated as random variables. Prior information about unknown parameters based on previously collected data and/or scientific knowledge can be combined with new data for parametric inference. For example, the fact that the parameter b in equation (1) takes the values 1.5 and 2.5 for rectangular and v-shaped sections, respectively, is an example of prior knowledge that can be used to form the prior distribution for one of the unknown parameters. Combination of the prior distributions and the model for the data results in the posterior distribution which can be used to obtain point estimates and interval estimates for the parameters. HSI has applied the Bayesian approach successfully to data on discharge and water level for discharge rating curve estimation.

In the section Models two statistical models for discharge and water level measurements are introduced. In the section Data a description of discharge and water level data is given. The two models are applied to the data in the section Results and a comparison between the models is made. Finally, in the last section conclusions are drawn.

MODELS

The Bayesian model for discharge rating curves currently used at HSI is given by

$$q_i = a(w_i - c)^b + \epsilon_i, \quad i = 1, ..., n$$

where n is the number of observations for a given site, (w_i, q_i) denotes the i-th pair of observations, ϵ_i is a mean zero measurement error such that

$$\epsilon_i \sim \mathrm{N}(0, \eta^2 (w_i - c)^{2b\psi}).$$

In essence this is the same model as the one presented by Petersen-Øverleir (2004). The parameter a is a function of φ and b, that is,

$$a = \exp(\alpha_0 + \alpha_1 b + \varphi)$$

where $\alpha_0 = 4.9468$ and $\alpha_1 = -5.3726$. This reparametrization is motivated by correlation between values of $\ln(\hat{a})$ and $\hat{b}$ which are based on data from HSI, and the values for α_0 and α_1 are selected such that there is no correlation between $\ln(\hat{a})$ and $\ln(\hat{a}) - \alpha_0 - \alpha_1 \hat{b}$, see Arnason (2005). The parameter ψ controls how the error variance behaves as a function of the expected value of q, and η^2 is a scaling parameter for the variance. This model will be referred to as Model 1.

Model 1 is not sufficient for about 5% of the data sets at HSI which calls for modifications. Here, a model is proposed that is an extension of

Model 1. It captures the main trend in discharge as a function of water level through the power-law part, $a(w-c)^b$, but a linear combination of B-splines is added, which allows for more flexibility than in Model 1. The form of this model is given by

$$q_i = a(w_i - c)^b + \sum_{l=1}^{L} \lambda_l B_{li} + \epsilon_i$$

where

$$B_{li} = B_l((w_i - w_{\text{low}})/r), \quad l = 1, ..., L, \quad i = 1, ..., n,$$

and the terms $B_l(z)$ are cubic B-splines (Wasserman, 2006) which have support on the interval $[0, 1]$, $r = w_{\text{upp}} - w_{\text{low}}$, w_{low} and w_{upp} are the lower and upper points of the interval influenced by the B-splines, respectively. Usually $w_{\text{low}} = w_{\text{min}}$, i.e., where w_{min} is the smallest observed water level. The quantity w_{upp} is selected as the 90th percentile of the water level observations or the number such that at least three water level observations are above it. The parameters in $\lambda = (\lambda_1, ..., \lambda_L)$, are unknown and L is the number of B-spline kernels. Further, the error terms are such that

$$\epsilon_i \sim \mathrm{N}\left(0, \eta^2 (w_i - c_2)^{2b_2}\right), \quad i = 1, ..., n,$$

where b_2 and c_2 are unknown parameters. Note that b_2 plays a similar role as ψ in Model 1. This model will be referred to as Model 2. Further, Model 2 is such that $\lambda_L = 0$ to avoid a jump at $w = w_{\text{upp}}$, and for $w < w_{\text{low}}$, $\mathrm{E}(q) = a(w-c)^b + \lambda_1$.

The Bayesian approach requires specification of prior distributions for each unknown parameter. The same normal prior distributions as used in Arnason (2005) are used here for φ, b and c, see details in Appendix. The prior distribution for b is a truncated normal distribution between 0.5 and 5. The posterior distribution of c will be influenced by its prior distribution but also by the smallest water level measurement, denoted by w_{min} since $c < w_{\text{min}}$. A normal Markov random field prior (Rue & Held, 2005) is assumed for λ, see details in Appendix.

The posterior distribution of $\theta = (\varphi, b, c, \eta^2, b_2, c_2, \lambda, \tau^2, \phi)$ given the data $q = (q_1, ..., q_n)$, $w = (w_1, ..., w_n)$ and w_{min}, is given by

$$\begin{aligned} p(\theta | q, w, w_{\text{min}}) &\propto \textstyle\prod_{i=1}^{n} p(q_i | \theta, w_i) \times p(\varphi) p(b) p(c) p(\eta^2) p(b_2) p(c_2) \\ &\times \; p(\lambda | \tau^2, \phi) p(\tau^2) p(\phi) \end{aligned}$$

where $p(q_i|\theta, w_i)$ is a normal density such that

$$p(q_i|\theta, w_i) = \mathrm{N}\left(q_i \middle| a(w_i - c)^b + \sum_{l=1}^{L} \lambda_l B_{li}, \eta^2 (w_i - c_2)^{2b_2}\right),$$

and the part $\prod_{i=1}^{n} p(q_i|\theta, w_i)$ is the likelihood function.

DATA

The data which are analyzed in this paper were collected by HSI water level measuring system and are from Norðurá in Borgarfjörður by Stekk. The river is located in the Western part of Iceland. The water level of Norðurá has been measured continuously since 1965. The data contain 35 pairs of discharge measurements (q), in m^3/sec, and water level measurements (w) in cm. Norðurá is a direct run stream with 500 km^2 drainage basin above Stekk. In direct run streams the discharge dependents heavily on the season and rainfall.

RESULTS

Here the two models introduced in the section Models are applied to the data from Norðurá in Borgarfjörður for comparison between the two models. Fig. 1 shows the fit of the two models to the data. There is a clear

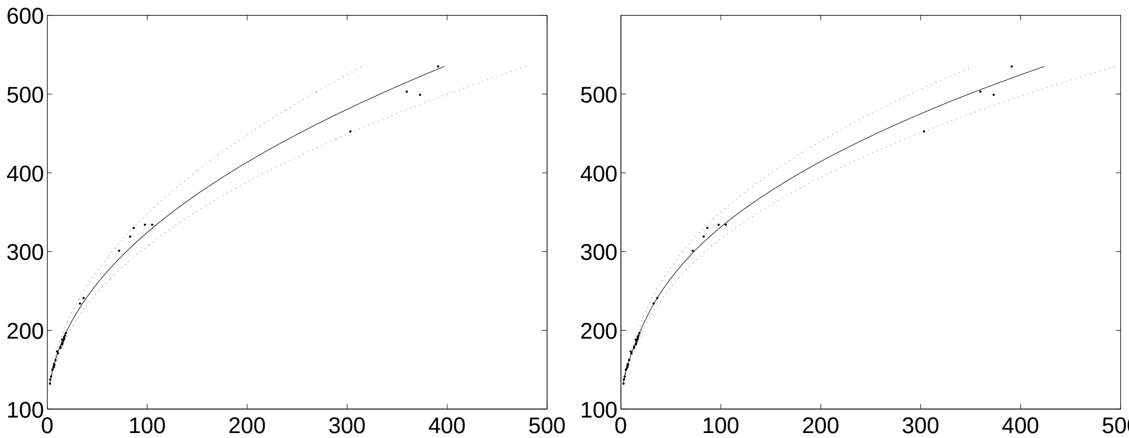


Figure 1: Water level is on the vertical axes (cm) while discharge is on the horizontal axes (m^3/sec). The points show the observed data from Norðurá in Borgarfjörður and the fit of Model 1 to these data (left panel), and the fit of Model 2 to the same data (right panel). The solid curves show the posterior median of E(q) while the dotted curves show prediction intervals.

difference between the two models. Both models fit the data very well for smaller values of water level while for larger values of water level Model 2 seems to perform better. This is due to the lack of flexibility of Model 1, it is not flexible enough to give a good fit to the few observations with large values of water level. The 95% prediction interval is wider for larger values of water level in Model 1 than in Model 2. This is mainly due to the fact that η^2, the parameter controlling the variance of the errors, is smaller in Model 2 than in Model 1. Fig. 2 shows the standardized residuals of the two models versus water level. In Fig. 2, it can be seen that Model 1 (left panel) is not flexible enough to handle the trend found in the standardized residuals while Model 2 yields more convincing standardized residuals. In general standardized residuals should not show any trend and appear to have the same variance for all values of the water level.

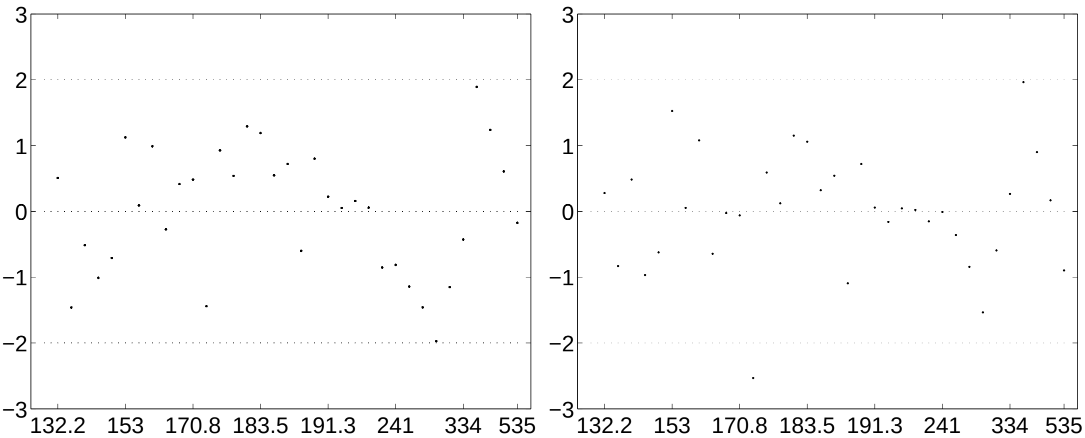


Figure 2: Water level is on the horizontal axes (cm) but the scale is nonlinear, standardized residuals are on the vertical axes. Standardized residuals for Model 1 (left panel) and Model 2 (right panel).

Fig. 3 shows the roles that the B-spline part (dotted line) and the standard power-law part (solid line) play in Model 2. The B-spline part picks up the extra trend in the data for the values of the water level below w_{upp} that the standard power-law part can not adjust for to the same extent by itself. This, in turn, allows the standard power-law part in Model 2 to give a better fit above w_{upp}. The B-spline part slowly dies out with increased water level and is practically zero above a value much smaller than w_{upp}. This behavior of the fit for Model 2 indicates that there is no breaking point in the discharge rating curve.

A model criterion called the deviance information criterion (DIC) (Spiegelhalter *et al.* , 2002) is used to further compare the two models. Three other

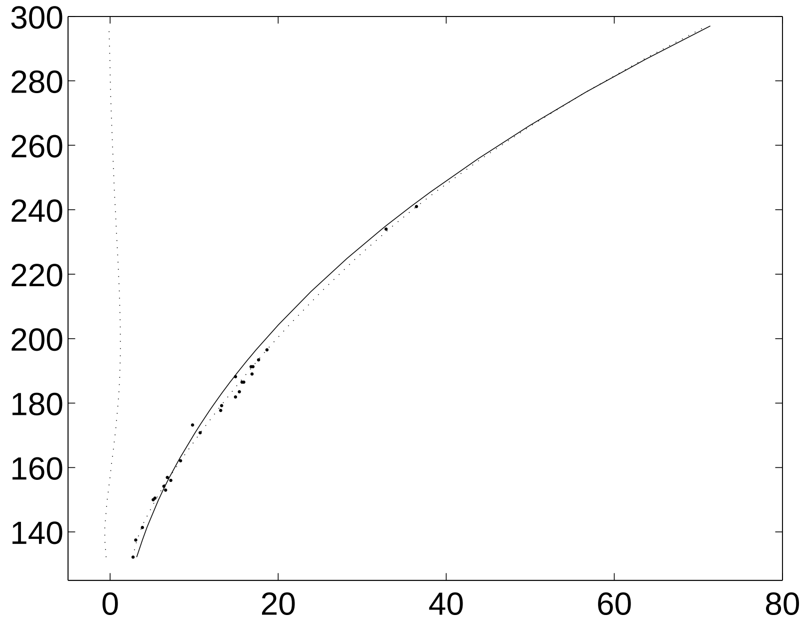


Figure 3: Water level is on the vertical axes (cm) while discharge is on the horizontal axes (m^3/sec). The standard power-law part (solid line) and the B-spline part (dotted line) of Model 2. The dotted line next to the solid line shows the sum of the two parts. The figure shows the water level values where B-splines have an effect on the model.

quantities are computed for each of the two models, namely, D_{avg} and $D_{\hat{\theta}}$ which are based on the likelihood function, and p_D, where $p_D = D_{\text{avg}} - D_{\hat{\theta}}$. The quantity p_D is the effective number of parameters. Further, DIC $= D_{\text{avg}} + p_D$, where DIC is such that the lower it is, the better is the fit of the model to the data. For details on DIC, D_{avg}, $D_{\hat{\theta}}$ and p_D, see Spiegelhalter *et al.* (2002) and Gelman *et al.* (2004).

The values of D_{avg}, $D_{\hat{\theta}}$, p_D and DIC for Models 1 and 2 are shown in Table 1. Here Model 2 has lower DIC than Model 1, the difference is more than twelve which is a substantial difference while a difference of size four or less leads to inconclusive results. This confirms that the added complexity of Model 2 does improve the fit. Model 1 has five effective parameters, however, in the simulation the estimate of p_D is 4.65 which is slightly different from five but this difference can be explained by the stochastic nature of the simulation. The estimated number of effective parameters in Model 2 is 7.66. The number of parameters in Model 2, if counted directly, is 23 since here $L = 15$, however, since the λ's are penalized through the prior of the λ's, the addition of c_2, λ, τ^2 and ϕ in Model 2 compared to the five parameters in Model 1 is equivalent to two or three unconstrained parameters.

Table 1: The deviance information criterion (DIC) and the effective number of parameters (p_D) for Model 1 and Model 2 along with D_{avg} and $D_{\hat{\theta}}$.

	D_{avg}	$D_{\hat{\theta}}$	p_D	DIC
Model 1	137.11	132.47	4.65	141.76
Model 2	120.95	113.29	7.66	128.62

Table 2 shows estimates of the parameters φ, b, c, η^2 and ψ in Model 1 while Table 3 shows estimates of the parameters φ, b, c, η^2, b_2 and c_2 in Model 2. The posterior mean of b is 2.17 for Model 1 while it is 2.51 for Model 2, so, the added flexibility of Model 2 yields a larger shape parameter. Yet, this increase in b will result in a large increase in discharge prediction for water level larger than w_{max}. The precision of these five parameters is better in Model 1 than in Model 2. For example, the 95% Bayesian confidence intervals for b and c are about three times and six times wider in Model 2 than in Model 1, see the 2.5 and 97.5 percentiles for b and c in Tables 2 and 3. The smaller precision seen in Model 2 results in less precision in the estimated discharge curve, $\mathrm{E}(q)$, than in Model 1, see Fig. 4. For the larger values of water level the width of the posterior interval for $\mathrm{E}(q)$ is around 35% greater in Model 2 when compared to Model 1. However, for water level values greater than 200 cm Model 1 appears to lack the curvature that the data suggest. The smaller precision in Model 2 is due to its complexity relative to Model 1 but what is gained is a better fit to the data.

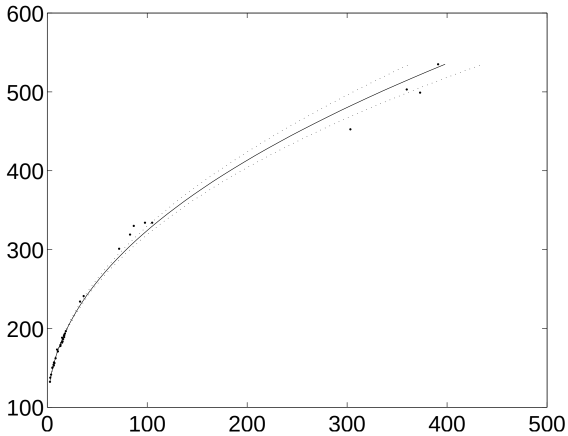
Table 2: Parameter estimates for Model 1.

	φ	b	c	ψ	η^2
Post. mean	-0.54	2.17	88.0	1.08	0.004
Post. median	-0.54	2.16	88.4	1.09	0.004
2.5 percentile	-0.66	2.03	78.7	0.94	0.002
25 percentile	-0.57	2.11	85.5	1.05	0.003
75 percentile	-0.51	2.21	91.0	1.13	0.005
97.5 percentile	-0.45	2.33	95.0	1.18	0.012

Table 3: Parameter estimates for Model 2.

	φ	b	c	b_2	c_2	η^2
Post. mean	-0.86	2.51	66.2	2.52	82.5	6.97e-009
Post. median	-0.88	2.53	63.0	2.55	82.4	4.01e-011
2.5 percentile	-1.38	2.05	24.6	1.96	51.1	4.42e-013
25 percentile	-1.05	2.35	50.1	2.34	71.7	5.30e-012
75 percentile	-0.66	2.68	81.5	2.71	93.5	4.58e-010
97.5 percentile	-0.30	2.94	118.0	2.94	113.0	4.45e-008

Posterior simulations for both Model 1 and Model 2 are stable and the simulated chains convergence in all in both cases. In case of Model 1 four chains of length 50 thousand are sufficient while for Model 2 four chains of length 100 thousand are needed to obtain adequate convergence.



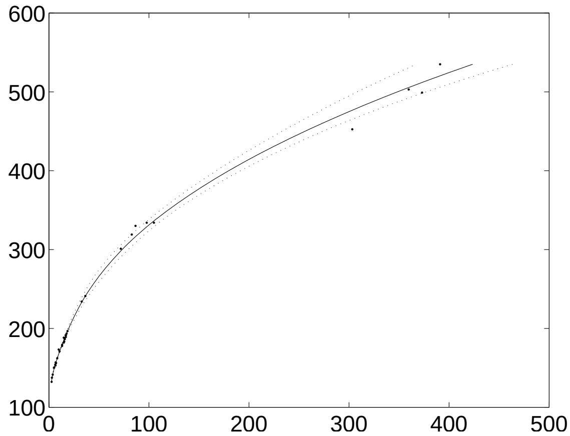


Figure 4: Water level is on the vertical axes (cm) while discharge is on the horizontal axes (m^3/sec). The points show the observed data from Norðurá in Borgarfjörður and the fit of Model 1 to these data (left panel), and the fit of Model 2 to the same data (right panel). The solid curves show the posterior median of $E(q)$ while the dotted curves show posterior intervals for $E(q)$ which are equivalent to confidence intervals in classical statistics.

CONCLUSIONS

Model 2 shows promising results when fitting rating curves in cases where Model 1 lacks the flexibility needed. The B-spline part of Model 2 is small relative to the standard power-law part but it catches the small deviation from the standard power-law model which results in a more convincing fit for Model 2 than Model 1. This is confirmed with DIC calculations where Model 2 yields a substantially lower value than Model 1.

Model 2 is formulated such that if Model 1 is the correct model or the adequate model then the B-spline part will be close to zero. However, further research is required to test the performance of Model 2 when Model 1 is the correct model.

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APPENDIX

The following prior distributions are proposed for the unknown parameters.

$$p(\varphi) = \mathrm{N}(\varphi|\mu_\varphi = 0, \sigma_\varphi^2 = 0.82^2)$$

$$p(b) \propto \mathrm{N}(b|\mu_b = 2.15, \sigma_b^2 = 0.75^2)I(0.5 < b < 5)$$
$$p(c) \propto \mathrm{N}(c|\mu_c = 75, \sigma_c^2 = 50^2)I(c < w_{\min})$$
$$p(\psi) \propto \mathrm{N}(\psi|\mu_\psi = 0.8, \sigma_\psi^2 = 0.25^2)I(0 < \psi < 1.2)$$
$$p(b_2) \propto \mathrm{N}(b_2|\mu_{b2} = 2.15, \sigma_{b2}^2 = 0.75^2)I(1 < b < 6)$$
$$p(c_2) \propto \mathrm{N}(c_2|\mu_{c2} = 75, \sigma_{c2}^2 = 50^2)I(c_2 < w_{\min})$$
$$p(\eta^2) \propto \text{Inv-}\chi^2(\eta^2|\nu_\eta = 10^{-12}, S_\eta^2 = 1)$$
$$p(\phi) = \mathrm{Beta}(\phi|\alpha_\phi = 1, \beta_\phi = 20)$$
$$p(\tau^2) \propto \text{Inv-}\chi^2(\tau^2|\nu_\tau = 10^{-12}, S_\tau^2 = 1)$$
$$p(\lambda|\tau^2, \phi) \propto \mathrm{N}(\lambda|0, \tau^2 D(I - \phi C)^{-1} M D)$$

where $I(A)$ is such that $I(A) = 1$ if A is true and $I(A) = 0$ otherwise. In the prior distribution for λ, I is an identity matrix, D and M are diagonal matrices and C is a neighborhood matrix with constants on the first off-diagonals, other elements are equal to zero.