ENCYCLOPEDIA OF EARTH SCIENCES SERIES

ENCYCLOPEDIA *of* SNOW, ICE AND GLACIERS

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Springer

Library of Congress Control Number: 2011922317

ISBN: 978-90-481-2641-5
This publication is available also as:
Electronic publication under ISBN 978-90-481-2642-2 and
Print and electronic bundle under ISBN 978-90-481-2643-9

Published by Springer
P.O. Box 17, 3300 AA Dordrecht, The Netherlands

Printed on acid-free paper

Cover illustration: photo 81332975 from Photos.com. © 2011 Photos.com

worldwide. Hydrologic simulation models that include snow are generally divided into three basic components, namely, snow cover, precipitation–runoff relationship, and runoff distribution and routing procedures. Most simulation models simulate the entire snow accumulation and melt season. Some of the models simulate only the snowmelt processes, whereas other models are used in conjunction with models of soil moisture accounting.

Snowmelt models that simulate runoff can be classified using a number of model characteristics. It begins with the division of the basin model into two major components, namely, a snowmelt model and a transformation model. The snowmelt model simulates the processes of snow accumulation and snowmelt, while the distribution and routing is accomplished by the transformation model which relates the snowmelt and rainfall-runoff volumes to discharge hydrographs. The most common transformation function is the unit hydrograph. For downstream channel reaches, the routing may be accomplished by a number of hydrologic routing techniques.

Singh and Singh (2001) have described snowmelt and ice melt modeling processes as well as concerned models in detail. Some of the important snowmelt runoff simulation models are listed below:

1. Snowmelt Model (SNOWMOD), India
2. Streamflow Synthesis and Reservoir Regulation (SSARR), USA
3. Snowmelt Runoff Model (SRM), Switzerland
4. University of British Columbia Watershed Model (UBC), Canada
5. Hydrologic Engineering Center (HEC-1) Model, USA
6. National Weather Service Snow Accumulation and Ablation Model, USA
7. HBV Model, Sweden

Summary

Modeling of melt runoff generated from snow and glaciers in a basin involves an understanding of snow accumulation, depletion, water percolation through snowpack, and routing of generated runoff up to the point of interest. In melt computation, mostly temperature index methods are used simply because of the availability of temperature data as compared to radiation data. Although there are a number of models available to simulate and forecast melt runoff, there is still much scope for improving modeling of melt runoff focusing on accurate estimation of spatial distribution of melt rate.

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Cross-references

Degree-Days
Surface Energy Balance

MELTING PROCESSES

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Definition

Melting refers to the change of state from solid to liquid, such as the change from ice to water. It occurs at 0°C for pure ice/snow at surface atmospheric pressures, although both changes in pressure and impurities (e.g., salt) can change the melting point. This review describes melting processes as they relate to ice and snow, with a focus on processes that contribute to both surface and basal melting on glaciers.

Surface melting

The main controls on surface melting of ice and snow are air temperature and solar radiation, although other factors such as surface albedo, rainfall, humidity, and wind can also be important (Hock, 2005). The importance of these factors has long been researched (e.g., Ångström, 1933), and is illustrated by discharge records from alpine glaciers, which show strong diurnal variability in the summer. Surface melting is usually the main way by which snow and ice bodies lose mass in temperate climates, although in cold climates (e.g., Antarctica) mass losses from iceberg calving and sublimation often dominate. In terms of surface melting, there are three main factors that control rates on ice and snow:

1. Net radiation

   The net radiation at the surface describes the difference between incoming and outgoing energy (Hock, 2005). Incoming energy is provided by:

   (a) Direct, diffuse, and reflected shortwave radiation that originates from the sun. On clear days, most shortwave radiation reaches the surface as direct sunshine, but on cloudy days diffuse radiation dominates due to the effectiveness of atmospheric water vapor, dust, and aerosols in scattering shortwave radiation (Paterson, 1994; Hock, 2005).

   (b) Longwave radiation emitted by objects such as clouds and surrounding mountain slopes after they have been heated, typically by incoming shortwave radiation.

   Outgoing energy occurs in the form of longwave radiation. Net melting occurs if there is less energy emitted than absorbed once the ice/snow has reached 0°C. Surface albedo is one of the most important factors in controlling how much shortwave radiation is absorbed by a surface and variations in albedo can produce dramatic differences in surface melt rates. For example, an old (firn) snow surface with an albedo of ~0.45 will melt approximately twice as fast as a new snow surface with an albedo of ~0.90 for the same amount of incoming solar radiation (albedo values from Paterson, 1994, Table 4.1).

In a study of melt patterns on a glacier on Ellesmere Island, Canada, Arendt (1999) found that the absorbed shortwave radiation flux accounts for 96–99% of the energy available for melt. In a summary of 16 studies by Hock (2005), it was similarly found that net radiation provides the majority of energy for melt, averaging 65.1%.

2. Sensible heat
Sensible heat refers to heat that is transferred between the atmosphere and an ice/snow surface, and is primarily supplied by warm air masses (Benn and Evans, 1998; Paterson, 1994). In general, it is less important than net radiation, with the review of Hock (2005) finding that it contributed an average of 26.4% of the total melt energy. However, the proportion of glacier melt that originates from sensible heat is very sensitive to wind, with the sensible flux sometimes dominating over net radiation on windy days (Holmgren, 1971). A rough ice surface also encourages wind turbulence and the transfer of sensible heat to the surface. Rain also produces a sensible heat flux, but is generally an insignificant factor for the annual surface energy balance of most glaciers (Hock, 2005). However, it can be locally important during heavy rain events on glaciers in temperate climates such as New Zealand, where Hay and Fitzharris (1988) found that it contributed 37% of the daily ablation during one prolonged event.
3. Latent heat
Energy from latent heat is produced from a change in state, and can warm an ice/snow surface when there is condensation of water vapor (e.g., dew) or the freezing of rainwater. It can also significantly raise temperatures where there is refreezing of percolating meltwater within a snowpack. It can be locally important, but in general it is the least significant source of energy for surface melt, with the review by Hock (2005) indicating that it accounted for an average of 7.8% of total melt energy.

Seasonal characteristics and snowpack processes

In the spring, melting and runoff from snow and ice bodies typically occur several days to weeks after air temperatures reach 0°C as energy is initially expended on warming the snow and ice to the melting point. Once melt begins, surface runoff from the lower (ablation) area of glaciers typically occurs via supraglacial streams as ice is essentially impermeable. These streams can enter crevasses and moulins, which enable the movement of meltwater to the interior or base of glaciers through englacial and subglacial streams. Direct measurements of melt rates on alpine glaciers record typical surface losses of ~50–100 mm day^{-1} in summer. For example, Purdie and Fitzharris (1999) recorded average summer surface melt rates of 96 mm day^{-1} on bare ice over the terminus of the Tasman Glacier, New Zealand, with a variation between 30 and 175 mm day^{-1}. However, surface debris had a significant insulating effect, with melt rates averaging only ~7 mm day^{-1} where the debris was 1.1 m thick.

In snowpacks and snow-covered parts of glaciers (mainly their accumulation areas), meltwater will not directly runoff the surface but will instead percolate vertically and refreeze where there are internal cold layers and irregularities in the snowpack. This creates internal ice lenses that can lead to retardation and vertical and horizontal channeling of meltwater. The percolation and refreezing process warms the snow via the release of latent heat, and is typically the dominant method by which snow in the accumulation area of many glaciers is warmed. Indeed, it is common for polythermal glaciers in locations such as Svalbard to have high altitude accumulation zones, which are warmer (i.e., typically at 0°C) than their low altitude ablation zones due to this process (Björnsson et al., 1996).

On polythermal glaciers, the refreezing of surface meltwater within and at the base of a snowpack can produce significant quantities of superimposed ice. In this situation, the melting process does not result in a net loss of mass, but the redistribution of it within the existing snow/ice body. Superimposed ice typically occurs at the boundary between the accumulation and ablation zones, and has the net effect of moving the equilibrium line altitude to a lower level.

Basal melting

Significant melting can occur at the base of glaciers and ice sheets if energy is available from sources such as geothermal heat, frictional heat due to ice flow, and sensible heat from meltwater that originated from the surface. The relative importance of these heat sources varies between ice masses, with surface-derived meltwater likely dominating on temperate glaciers (due to the predominance of crevasses and moulins), and geothermal heat dominating in the center of the large ice sheets (where there is little basal motion and no surface meltwater). The inaccessibility of the glacier bed makes it difficult to directly measure the importance of these processes, although the recent discovery of a large number of subglacial lakes beneath the Antarctic Ice Sheet (Siegert et al., 2005) indicates that basal meltwater can persist for a long time once formed.

The influence of pressure on the melting point can be a significant factor for thick ice masses, depressing it at a rate of 0.072°C per million Pascal (MPa) (Benn and Evans, 1998). A 1 km thick glacier has a basal pressure of ~8.8 MPa, which means that the pressure melting point at the base of the ~4 km thick Antarctic Ice Sheet is approximately −2.5°C. The change in pressure melting point with depth means that a body of ice that is solid near the surface can melt at its base without any change in temperature. In addition, if all ice is at the pressure melting point (e.g., in a temperate glacier) then local variations in pressure can be expected to cause local melting. This regelation process can result in melting on the upstream side of bedrock bumps where pressures are higher than average, and refreezing on the downstream side where pressures are reduced. This explains how water can be produced at the glacier bed without a change in temperature.

Summary

In summary, surface melt processes on ice and snow are strongly tied to climatic conditions, with net radiation typically providing the main source of energy for melt. Internal factors such as geothermal heat fluxes are more important for determining basal melt rates and processes, although climate conditions are also important where surface meltwater is able to reach the bed of a glacier via moulins or crevasses.

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Cross-references

Albedo
Bottom Melting or Undermelt (Ice Shelf)
Calving Glaciers
Equilibrium-Line Altitude (ELA)
Refreezing of Meltwater
Subglacial Drainage System
Sublimation from Snow and Ice
Surface Energy Balance

MELTWATER CHANNELS

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Definition

Channels eroded by glacial meltwater into bedrock or sediment beneath or near the margins of glaciers and other ice masses.

Introduction

Meltwater channels encompass a wide variety of erosional channel features formed by flowing water sourced from melting glacier ice. They form either beneath the ice (subglacial), along the margins of the ice (ice marginal), or immediately in front of the ice (proglacial). Meltwater channels may be single, discrete channels or branching networks and range in size from small (meter scale) channels (e.g., Hambrey, 1994) to large-scale tunnel channels, hundreds of meters deep and kilometers wide (e.g., Jorgensen and Sandersen, 2006). However, most references to meltwater channels refer to channels that range from meters to tens of meters width and depth and range from tens of meters to kilometers in length.

Meltwater channels are one of the most obvious features of past and present glacial environments and provide a complex but powerful means of delineating past ice limits, glaciological conditions, and deglaciation trends for glaciers and ice sheets. The majority of studies regarding meltwater channels have focused on the Pleistocene ice sheets of the Northern Hemisphere. For example, the recognition of different channel types, combined with mapping of channel distribution have allowed detailed meltwater flow patterns and deglaciation histories to be constructed for the Pleistocene British Ice Sheet (e.g., Glasser and Sambrook Smith, 1999; Evans et al., 2005; Greenwood et al., 2007), Fennoscandian Ice Sheet (e.g., Kleman et al., 1997), and Laurentide Ice Sheet (e.g., Dyke, 1993). However, meltwater channels have been described from many other regions and from various intervals of earth history. For example, Lewis et al. (2006) described meltwater channels formed by subglacial drainage from beneath the East Antarctic Ice Sheet during the Miocene and Williams (2005) reported channels thought to have formed by subglacial meltwater 1.8 billion years ago in what is now the arid northwest of Australia.

Temperature, glacier thermal regime, and meltwater

The amount of meltwater generated in a glacial system varies according to the regional climate and associated thermal regime of glaciers (Eyles, 2006). The thermal