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Abstract

The hydrology of glaciers concerns the transfer of meltwater through an ice mass. Conceptually, a model for the hydrology of mid-latitude glaciers provides a broad basis for our current understanding. In general, a glacier's hydrology is divided into three core environments: the surface, interior, and ice-bed interface. The structure and development of drainage within these environments controls the efficiency and manner in which meltwater is delivered to streams emanating from a glacier. Compared to flow over exposed bare ice, snow significantly delays the rate at which meltwater is transferred over a glacier surface. Micro- and macro-scale drainage can enable water to drain through a glacier's interior. Meltwater at the ice-bed interface follows slow or fast flowpaths dependent on water volumes and the nature of the glacier bed. The progression of the melt season increases the efficiency of drainage through a glacier, consequently influencing both the volumes and quality of runoff.

Main Text

Introduction

Glacier hydrology is the scientific study of the storage and transit of liquid water associated with snow and glacier ice. This includes water present at the surface of, in the interior of, at the basal interface of and emerging from, bodies of snow and ice. Today, glacier ice covers around 11% of Earth's surface area, accounts for ~70% the planet's freshwater resources, and underpins the water security of over a billion people. For millennia, glacier-fed rivers have been critical for irrigation and agriculture, but have always posed a threat of flooding. Meltwater runoff from glacierized catchments is an important resource for hydroelectric power, accounting for 50% or more of the electricity produced in mountainous countries such New Zealand, Peru, Pakistan, Switzerland, and Norway. Recently, focus has turned to sporadic, catastrophic and destructive releases of meltwater from glacierized terrain, the mechanisms for, and prediction of which remain poorly constrained. Therefore, the hydrology of glaciers is an important, topical subject demanding continued investigation, particularly in the context of the recent trends observed in glacier extents and the Earth's atmosphere-climate system.

Glaciers represent a frozen environmental asset: natural reservoirs storing freshwater over a wide range of time-scales, from hours to millennia. Long-term freshwater storage, at time-scales greater than a year, typically involves glacial ice; intermediate and shorter-term storage more usually involves seasonal snow and liquid water and is best exemplified by the seasonality of glacier runoff. The hydrology and runoff from glaciers is fundamentally controlled by the surface energy balance (see **Hock, 2005**). For the vast majority of Earth's glaciers, outside the tropics, mass is accumulated as snow during the winter, when the energy balance is at a minimum due to low air temperatures and reduced solar radiation receipt. During summer months, increased air temperatures and solar irradiance provide a highly positive surface energy balance, resulting in the melting of snow and glacier ice (termed *ablation*). Thus, glaciers modulate streamflow by releasing the most runoff during the warmest, driest periods in the year when all other sources of water are minimal. The volume of meltwater runoff that a glacier provides is a function of its surface area and ablation rate. The melt process ensures runoff from glacierized catchments correlates well with air temperature and exhibits a strong diurnal signal in the ablation season *hydrograph*. Consequently, glaciers exert a strong influence over a catchment's runoff, even if glacier ice only occupies a small proportion of the catchment area.

Within a glacierized catchment, the drainage system that routes meltwater from the glacier surface to the ice margin can be conceptualized as a cascade (**Figure 1**). Meltwater can travel along flowpaths at the surface (*supraglacial*), within the glacier's interior (*englacial*), and at the ice-bedrock interface (*subglacial*). The structure, function and interaction of these systems determines the efficiency and transit time by which meltwater travels from the point of melting to the glacier terminus. Accordingly, runoff from a glacier is controlled by the characteristics and hydraulic properties of these three drainage system components which modulate both meltwater volumes and water quality through solute acquisition and sediment evacuation. Hydrological coupling between the supra-, en- and sub-glacial environments is fundamental in mediating the manner in which a glacier responds to environmental variability, primarily due to the role water plays in controlling ice motion. Here, focus is on mid-latitude, temperate valley glaciers, but for details on the hydrology of non-temperate glaciers and ice sheets, see **Irvine-Fynn et al. (2011)** and **Chu (2013)**, respectively.

[Figure 1 here]

Supraglacial hydrology

The character of the supraglacial environment can be categorized into three key states, each with their own distinctive hydrology: snowpack, *firn*, and bare ice.

The snowpack

The hydrology of snow in the supraglacial environment is generally the same as for other terrestrial snowpacks (see **DeWalle and Rango, 2008**). Snowpacks on glaciers are the result of multiple snowfall events; in mid- and high-latitudes, these typically happen during winter months when little or no melt occurs, while for tropical glaciers, snow accumulation transpires during warmer summer months. Consequently, glacier snowpacks are highly heterogeneous, exhibiting variations in crystal grain size and density from each snowfall event, leading to a complex, layered stratigraphy. After deposition, the snow structure is transformed by the pressure of overlying snow, by wind redistribution and compaction, and by internal processes of melting, refreezing, sublimation and condensation. Such *snowpack metamorphosis* is driven by temperature gradients and snowpack age, and typically enlarges the crystal size or joins grains together (*sintering*), further influencing snowpack density. Combined, these processes result in local variations in snow density, ranging from 50 to 400 kg m^{-3}, and lead to contrasts in snowpack porosity and hydraulic conductivity, not only with depth but also over space.

As the melt season commences, snow surface meltwater begins to percolate into and *warm* the snowpack. The infiltration of liquid water and the latent heat released by refreezing begins to overcome the *cold content* of the snow, raising the snowpack temperature towards the melting point. Densification of the snowpack occurs as diurnal melt-freeze cycles cause the refreezing of percolating meltwater either within the pore spaces of the snow or as more massive ice lenses. Continued melt *ripens* the snowpack as gravity-driven percolation causes the wetting front to descend away from the surface, with colder snow below. The percolation rate of meltwater can be approximated by Darcy's Law, with velocities of 10^{-4} to 10^{-5} m s^{-1}. Critically, due to the structure and density variations within the snowpack, the wetting front is neither uniform nor horizontal, and preferential flowpaths or *flow fingers* develop. The velocity of meltwater through a ripe snowpack is dependent on the physical structure of the snow and the water flux: greater volumes of melt travel more rapidly. As the snowpack becomes isothermal (at melting point) the increased water content raises the homogeneity and hydraulic permeability. Meltwater content and flow will be highest (and most concentrated) where melt rates are greatest. Drainage through a saturated snowpack is driven by the local slope and transport velocities can increase to 10^{-2} m s^{-1}.

Over the accumulation area of a glacier, at the apex of the summer melt cycle, the varied physical and thermal properties of the snowpack may be classified as three distinct *snow facies*: (i) *dry snow* is typically only found at very high elevations (or in the interior of ice sheets), where no melting occurs, the water content remains at 0%, and the snow makes no contribution to runoff; (ii) *percolation snow* occurs at intermediate elevations where some summer melting occurs, meltwater (~3% water content by volume) penetrates the snowpack or drains into the firn (see *The firn*), but refreezing and storage occurs ensuring melt does not contribute significantly to runoff; (iii) *wet snow* arises where summer seasonal melt raises the snowpack to 0°C and vertical percolation results in saturated snow (8 to 15% water content) and lateral flow occurs particularly at the glacier ice interface, or where ice lenses or layers occur. In contrast to cold or polythermal glaciers, the summer melt season warms the entire snowpack on temperate glaciers and the dry and percolation snow facies are not observed.

At lower elevations, in the ablation zone, the increasingly thin and saturated snowpack may become *slush* (>15% water content). Slush flows themselves may contribute to the removal of snow from the glacier surface. The slush zones commonly follow the seasonal retreat of the transient snowline, and in this zone supraglacial lakes may form where glacier surface gradients are low and/or topographic depressions exist. Undulations in glacier surfaces can be formed by ice flow itself and/or by the topography of the glacier bed. Meltwater may also pond in crevasses during the early part of the melt season. Lake volumes of 10^4 m^3 have been described on valley glaciers, while on the Greenland Ice Sheet, supraglacial meltwater lake volumes of 10^7 m^3, with extents of up to ~9 km^2 have been reported. Supraglacial lakes are transient features usually forming early in melt season as supraglacial drainage is initiated, when saturated snow- or slush-plugs prevent flow through supraglacial stream courses. However, on ice sheets, lakes may persist between melt seasons.

At the end of the summer, the snowline position broadly corresponds to the *equilibrium line altitude* (ELA), which is the average elevation of the area on the glacier where accumulation equals ablation over a 1-year period. At the ELA, slush and supraglacial lakes may persist during summer, but superimposed ice is also evident. Superimposed ice forms where meltwater refreezes at the glacier ice interface, most typically in early summer as meltwater percolates through the snowpack, but also in early autumn when residual melt draining from the accumulation area and precipitation refreeze as air temperatures decline. Although superimposed ice can form across the entire ablation zone, the high ablation rate at lower elevations removes evidence of its formation except at the ELA where only the winter snowpack is melted. Superimposed ice can reduce the permeability of surface glacier ice so that accumulation of meltwater on the glacier surface is pronounced during the early period of the ablation season.

The firn

Firn represents the transition between snow and glacier ice, and is formed through the processes of *firnification*. This process represents the compaction and reorganization of snow and ice crystal structure, which in turn increases its density to >400 kg m^{-3}. Snow that survives over a full annual cycle becomes firn. The rate at which this snow transforms into firn (and ultimately into glacier ice) is dependent on the accumulation rate of snow, the temperature of the snow, and the presence of meltwater in the evolving snowpack. These controls impose a gradual reduction in the air or void space within the snowpack, and firn becomes glacier ice upon the occlusion of intercrystalline air passages to bubbles, at a density of ~830 kg m^{-3}. The density of firn increases with vertical distance from the snow surface, and also with proximity to the ELA where production and refreezing of meltwater is commonplace. These spatial trends towards an increasingly tight packed crystal structure results in a porosity gradient. Characteristically, firn is found above the ELA, and its hydrology is important for runoff sourced in a glacier's accumulation area.

Due to the presence of firn, the accumulation area of a glacier represents a *confined aquifer*. Meltwater derived from the surface snow can percolate through the snowpack and into the firn layers. The meltwater drains through the higher porosity, unsaturated firn to form a saturated firn layer at depth where the reduced permeability impedes further drainage. This saturated firn layer progressively develops following the onset of the melt season; the volume of water retained in this aquifer is dependent on, amongst other things, the snow melt rate.

The saturated firn layer on valley glaciers has been reported at depths of up to 15 m below the firn surface, with thicknesses ranging from 1 to 7 m above the firn-ice transition. The meltwater transit velocities of 10^{-4} to 10^{-5} m s^{-1} in the saturated layer emphasize the firn's ability to delay snowmelt runoff at time-scales of days to weeks. The slowed transport of meltwater in accumulation area firn reservoirs results in the damping of any diurnal melt signal during summer. However, as the melt season ends, the saturated firn drains down-glacier or meltwater refreezes in situ (aiding firnification), at least partially reducing the firn aquifer's liquid water content.

The bare ice

During the melt season, the decrease in snowpack extent, and up-glacier recession of the transient snowline towards the ELA, results in exposure of glacier ice and initiate rapid supraglacial runoff from the ablation area. If unimpeded, this meltwater flows across bare surface ice at much higher velocities than those seen in the snowpack or firn.

On all glaciers, during the melt season, the ablation area represents a *transient thermal layer*: in winter, the glacier ice surface is cooled below the pressure melting point (PMP), while during spring and summer the increasingly positive energy balance driving ablation ensures the seasonally cold surface layer initially approaches and then remains at the PMP. Because glacier ice is not a pure medium, the energy balance that drives melt is not uniform over space. Preferential radiation absorption at crystal boundaries or the presence of dust and/or impurities on the ice surface generates micro- (micrometer) to small- (centimeter) scale topographic variations. Meltwater may percolate through or along these flowpaths and, by increasing flow volumes, create rills on the ice surface.

Incident solar radiation also penetrates the upper-most glacier ice, typically to a depth of ~2 m. Through melting at subsurface crystal boundaries, the surface ice density can be reduced to ~550 kg m^{-3} creating a porous *weathering crust* through which water can be transported. Observations have suggested this shallow weathering crust zone behaves like a perched aquifer, with the capacity to store and release meltwater; however, this weathering crust aquifer is typically saturated, with its water table just a few centimeters below the ice surface. Local differences in ice structure (e.g. foliation) result in contrasts in hydraulic permeability in the ice, which may contribute to the formation of preferential flowpaths within the weathering crust ice. Transport rates in the weathering crust are typically between 10^{-4} and 10^{-6} m s^{-1}.

By following the steepest surface gradient, surface rills coalesce to form supraglacial stream channels. Akin to terrestrial river catchments, supraglacial stream channels commonly develop dendritic or arborescent drainage patterns, with drainage densities ranging from <10 to >20 km km^{-2} which generally decrease up-glacier where ablation is reduced. Due to the convexity of glacier ablation areas, elongated, subparallel drainage patterns are often observed. Similarly, the structural control imposed by down-glacier longitudinal foliation and annealed crevasse traces can result in rectilinear surface drainage systems. Delays of between 1 and 11 hours have been observed between peak melt production and peak supraglacial stream discharge; these delays relate to catchment size and geometry, but also the hydraulic properties of the weathering crust. The form of supraglacial flowpaths and stream catchment areas varies glacier to glacier: drainage networks may be disrupted by crevasses (see ***Englacial hydrology***) occurring in response to site-specific glacier motion and the associated stresses in the ice.

The geometry of supraglacial streams is controlled by, amongst other things, stream catchment area and ablation rate. To accommodate increasing meltwater discharge, supraglacial stream flow velocity and water depth increases while channel width remains relatively constant. Flow velocities of >1 m s^{-1} are common, even in small streams. Channel roughness is very low; the dimensionless Manning's 'n' for supraglacial channels is typically <0.03. As is seen in terrestrial river systems, both meandering and step-pool morphologies can develop in supraglacial streams. Meander wavelength (with sinuosity of 1.0 – 2.5) relates to discharge and stream width, while the sinuosity of meanders is inversely associated with reach-length channel slope. Step-pool sequences appear to develop where channel slope exceeds ~15°.

The incision rate of supraglacial streams is principally related to channel slope and discharge. Viscous fluid friction within the turbulent flow can accentuate channel incision. Supraglacial stream water equilibrates to around +0.1°C, meaning there is also sufficient thermal energy available to induce downcutting of ice-walled channels at rates of ~2 mm hr^{-1}. On glaciers in temperate latitudes, typically the ice surface ablation broadly matches stream incision rates (~0.1 m day^{-1}), and streams reform each melt season. Conversely, in colder climates, stream incision may be an order of magnitude greater than glacier ablation, resulting in the development of progressively deeply incised 'canyons' >2 m deep and persistent over many years. The process of meandering will continue at depth, and the planform of the surface expression of the supraglacial stream may not reflect the active channel.

As a result of the low permeability of surface and near-surface ice, the ablation area of a glacier exhibits a rapid hydrological response to changing melt rates. The rapid transport of meltwater via supraglacial streams in the ablation area ensures that the diurnal melt signal is apparent in the meltwater discharge delivered from the bare-ice of glaciers. However, the configuration of drainage structure in the supraglacial environment is, in part, dependent on the occurrence of features which provide access to the englacial environment.

Englacial hydrology

Massive glacier ice is permeable, allowing englacial water to permeate through it slowly. With a density of 830 to 910 kg m^{-3}, at the microscale, glacier ice crystals are surrounded by a film of liquid water, and these films form a network of *veins* at the junction between three or more individual ice crystals. Typically 10^{-1} to 10^{1} µm in diameter, the size of the veins depends on the ice crystal size, meltwater solute content, pressure and temperature. Ice crystals range from <1 mm to 10 cm in diameter, and the size of the vein junction is positively correlated to the crystal size. The process of ice crystal formation itself rejects solutes and impurities, resulting in solute-rich (or saline) water within the intercrystalline or *interstitial* vein network. The lower freezing point of saline water compared to that of pure water allows for the existence of larger veins. Increases in overburden pressure reduce the PMP of ice, potentially enlarging the vein network in ice nearer the glacier bed; the hydraulic permeability of the ice in the lowermost several tens of meters in a glacier increases. Englacial strain heating resulting from the flow of the ice can also contribute to interstitial meltwater (0.01% by volume). However, irrespective of crystal size and solute content, as ice temperatures decrease, the vein network contracts; the porosity and permeability of ice at the PMP is considerably higher than for cold ice. Liquid water can be present in ice substantially below the PMP, but usually <0.1% by volume, while the water content of glacier ice at the PMP may rise to several percent by volume. The flow in the vein network is small, however, with laboratory studies suggesting velocities of between 10^{-8} and 10^{-13} m s^{-1}. The presence of air bubbles and debris, capillary forces, the deformation and recrystallization of ice, as well as high confining pressures, are likely to contribute to such slow flow rates.

Despite the low rate of water transport, the interstitial vein network is prevalent throughout a glacier, and water will generally flow towards the glacier bed. Assuming ice is a porous medium, water moves from high to low *hydraulic potential* (Φ). Hydraulic potential is the sum of the gravity-defined potential energy and the water pressure defined by the force imposed by the overlying ice. Theoretically, water flows perpendicular to the 3-dimensional plane with equal hydraulic potential – the *equipotential surface* – and follows the steepest potential gradient. On valley glaciers, equipotentials typically dip up-glacier at 11-times the ice surface slope. Using calculations based on the glacier geometry (e.g. ice extent, topography and thickness) it is possible to calculate an informative overview of likely flowpaths in the englacial environment.

Water flow, directed by the hydraulic potential, generates heat proportional to the water flux, further enlarging already large veins in the ice. As such a vein enlarges, its water pressure drops compared to smaller veins. Water flow will be directed towards the larger evolving conduit, and consequently, large conduits develop and grow at the expense of smaller flowpaths.

This results in the development of an arborescent drainage pattern following the gradient of equipotentials through the englacial environment. However, studies of flowpaths in the englacial environment from direct observations (*glacier speleology*), ice-penetrating radar, and video imagery from boreholes do not appear to fully support the theoretical englacial drainage system as driven by hydraulic potential. Englacial conduits following sub-horizontal pathways, vertical descent depths of up to 60 m interrupted by small steps and horizontal galleries, or small (<0.1 m) englacial drainage structures with shallow dipping orientations have contributed to a growing realization that structural controls (*foliations*, fractures, variations in ice rheology etc.) may provide a greater influence on the geometry of drainage features, and local hydraulic potential fields, than was previously appreciated.

Crucially, in view of the limited water transfer capable by the *primary permeability* of englacial ice itself, it is the larger-scale structures (e.g. *crevasses*, fractures and *moulins*) that provide the pathways for the bulk of seasonal meltwater to access the englacial environment (**Figure 2**), even if hydraulic potential governs the general direction of water flow. These macroscale flowpaths that represent *secondary permeability* may allow transit velocities of between 10^{-1} to 1 m s^{-1}. Most commonly, crevasses exist as direct pathways to the glacier interior. Such crevasses form in response to high tensile strain rates, and have depths of up to ~30 m in temperate ice (ice at the PMP), and greater in colder, more brittle ice. Crevasses truncate any supraglacial streams they intersect and sequester water from them. Energy dissipation by flowing water at the crevasse base may develop a vertically descending conduit at depth. The stress field within the ice causes a lateral deviation of such evolving conduits, with observations suggesting a down-glacier orientation of ~25° from vertical. Where ice flow causes crevasses to close, but supraglacial drainage continues to incise a notch into the up-glacier edge of the former crevasse, a vertical shaft (*moulin*) forms. Moulins from <1 m to 10 m in diameter may be long-lived features, persisting for multiple melt seasons.

[Figure 2 here]

Where crevasses or moulins become water-filled, the process of hydrofracture may arise. Here, the mass of meltwater promotes fracture extension at the base of the crevasse or moulin. The depth of the hydrofracture depends on the volume of stored water and/or the persistence of meltwater delivery to the fracture locus. In the absence of a continuous water flux to the fracture base, refreezing will reduce the likelihood of the fracture descending to greater depths. Under optimal circumstances, crevasse hydrofracture has been observed of glaciers and ice sheets to propagate to depths of 10^2 to 10^3 m in short (hours - days) time-scales. As with crevasse and moulin drainage path formation, the changes in effective stress field with increasing depth can result in the hydrofracture flowpath becoming inclined at progressively shallower angles. Hydrofracture is also not limited to a vertical orientation, and appropriate water pressures and/or structural weaknesses can facilitate shallowly-dipping drainage path formation.

It is possible for supraglacial streams themselves to develop into englacial conduits, especially where stream incision exceeds general ice ablation rates: snow bridges and metamorphosed and deforming ice can cover these incised flowpaths, resulting in enclosed, high-roofed englacial channels called *cut-and-closure channels*. The geometry of these channels resemble supraglacial streams, with flow generally being at atmospheric pressure. Processes of vertical incision and knick-point retreat rates of 0.1 to 0.5 m day^{-1} in these channels can lead to the formation of moulin-like forms at the channel head. Similarly, sub-horizontal channel forms have been suggested to arise from the entombment of conduits formed at the base of the saturated snow zone in the accumulation area of ice masses. Crevasses may also lead to sub-horizontal drainage structures: water flow in crevasse bottoms may initiate a drainage route, and subsequent crevasse closure can isolate the drainage structure from the supraglacial environment. Speleological investigations have also revealed that ice structures - such as refrozen or annealed crevasse and/or fracture traces, debris bands, thrust faults, and highly permeable ice - all provide locations of high hydraulic transmissivity, supporting conduit inception. Studies have reported englacial fractures can be found throughout 96% of a glacier's depth. Fractures in ice can also result from frictional drag at the glacier bed, with basal crevasses propagating vertically upwards into the englacial environment.

Current wisdom suggests englacial drainage flowpaths close up over winter months and reopen each summer, when water pressures decrease with channel enlargement. The closure rate of water-free ice-walled channels is a function of ice overburden pressure and viscosity, largely controlled by temperature. In cold ice, even small channels can persist for over a year. In contrast, temperate ice deforms rapidly and, once empty of water, conduits close rapidly: theoretically, at a depth of 100 m, a conduit's diameter may halve in <20 days. However, englacial conduit closure rates are substantially reduced if they are water filled, as the stored water impedes ice deformation. The englacial system, therefore, can represent a reservoir with capacity to store considerable volumes of water, particularly over winter months. Empirical water balance studies have demonstrated disparities between input and output meltwater volumes that suggest such processes of storage and release occur within the englacial hydrological system.

Subglacial hydrology

Assuming the glacier bed is at the PMP, water reaching the ice-bed interface flows towards the glacier terminus, orthogonal to the equipotential contours formed by the intersection of the equipotential surfaces with the bed. Water can exist at the ice-rock interface beneath cold ice areas due to the hypersalinity of mineralized waters resulting from solute rejection during melting and refreezing. Meltwater at the glacier bed may also be sourced from basal and frictional heating in areas of temperate ice, although annually this rarely exceeds ~10^{-1} m of melt. Once present at the ice-bed interface though, meltwater flow is complicated by variability in both supply and bed conditions.

Where ice is thin, channel enlargement by viscous heat dissipation may offset channel closure, and thus channels would be open with meltwater flowing at atmospheric pressure. In this situation, bed geometry is the primary factor in defining the orientation of subglacial channels. However, observed channel locations have often appeared to mirror those reconstructed from hydraulic potential assuming water at ice-overburden pressures. To reconcile this dichotomy, it is quite possible that drainage pathways close by deformation in winter months and subsequently reopen each summer. Since conduits are water-filled, and therefore at (or temporarily above) ice-overburden pressure during re-formation in the spring, their initial location is controlled by hydraulic equipotential pressures. Subsequently, and despite reduced water pressures, melt rate is insufficient to allow lateral migration to the location dictated by theoretical open-channel flow (i.e. the low point of the glacier bed). Nonetheless, there is a broad agreement that water flow at the bed occurs in one or both of two qualitatively and hydraulically different conceptual flow systems: *distributed* or ‘slow’ flow and *discrete* or ‘fast’ flow.

Distributed (slow) subglacial drainage

Distributed subglacial drainage occurs through spatially extensive, non-arborescent (or anastomosing) flow pathways at relatively slow velocities of $<10^{-2}$ to 10^{-3} m s^{-1}. For a glacier with a hard bedrock sole at the PMP, micro-scale pressure-driven melting and refreezing (*regelation*) occurs around bedrock protuberances up to ~ 1 m long. The meltwater generated by this process flows around bedrock bumps and clasts as a mm-thick *film*. Sedimentological analyses reasoned that the lack of particles <0.2 mm at the rock-ice interface demonstrated the presence of water films capable of eroding finer particles, while numerical analyses indicated an upper limit to film stability of ~4 mm. Raised viscous heat dissipation in thicker films, or local increases in water discharge due to supply or bedrock variability, result in the development of preferential drainage paths and protochannels. Where ice flow is sufficiently fast, ice may separate from the bed in the lee of a bedrock protuberance, leading to the development of cavities. Subglacial meltwater may collect in, and thermally enlarge, cavities. The initial cavity size is dependent on bed topography, ice velocity and basal shear stresses. Cavities may be either autonomous or interconnected by narrow channels (*orifices*) where water flow is sufficient. A network of interlinked cavities forms an anabranching *linked cavity system*. For small discharges, ice melt is unimportant for maintenance of the cavity system, so small increases in water pressure lead to a greater carrying capacity of the system. However, if meltwater discharge through the inefficient linked cavity system increases further, preferential channel-like flowpaths can develop.

Many glaciers are not entirely in direct contact with consolidated bedrock. Instead, subglacial investigations and observations of recently deglaciated surfaces commonly indicate the presence of unconsolidated sedimentary deposits. With a hydraulic conductivity typically in the range 10^{-6} to 10^{-12} m s^{-1}, *Darcian porewater flow* is likely through subglacial sediments, although extremely difficult to measure directly. However, macroporous or preferential flowpaths may develop in subglacial till, and porewater flow is likely to occur in conjunction with other more efficient drainage systems.

Discrete (fast) subglacial drainage

Discrete subglacial drainage occurs through channels or conduits, typically forming a stable arborescent network along which meltwater flows rapidly at velocities of 10^{-1} to 10^{-2} m s^{-1}. The relative instability of distributed drainage systems to increased water flux means that they break down as water discharge accumulates either gradually down-flow or locally due to point inputs. Here, preferential drainage paths develop by melting the overlying ice to form semi-circular subglacial channels, *Röthlisberger*- or *R-channels*, which follow the hydraulic gradient. However, water pressures measured in subglacial channels have been found to be higher than those calculated for ideal R-channels, indicating unusually rough or sinuous conduits, or a cross-sectional shape that is broad and low rather than semi-circular. In these *Hooke*- or *H-channels* melting is concentrated on the conduit walls since channels are not continually water-filled, and lateral closure is limited by friction with the bedrock. Where bedrock may be readily eroded or where major subglacial channels persist for sufficient time, such channels may incise into the rock rather than the ice, forming *Nye*- or *N-channels*.

In the case of a poorly-consolidated sediment bed, due to the instability of films and porewater flow to externally driven variations in water flux, it is thought that shallow *anabranching canals* incise into the deformable bed material. Canals are enlarged by water flow removing sediments, with sediment creep and failure of the canal wall counteracting the tendency for growth.

Given the spatially restricted nature of discrete subglacial drainage and the spatially extensive nature of distributed subglacial drainage, it is anticipated that the two systems likely coexist side-by-side. Discrete channels are likely to form at point sources such as beneath moulin delivery points, and to extend down-flow of them, irrespective of their location. Channels will also generally increase in representation down glacier, where accumulated subglacial meltwater flux is largest. In between these channels, distributed drainage will dominate. Indeed, at temperate glacier beds at the PMP, the distributed component will be ubiquitous between discrete channel pathways. Critically, evidence suggests not only that such a combination of subglacial drainage systems do coexist, but that they interact at a variety of scales. For example, coupling between porewater flow and channelized flow has been observed whereby high daytime water pressures within channels drives water from them into the surrounding distributed system. At night, when the channel water pressures are markedly lower, that water returns, transferring solute and suspended sediment from the surrounding locality to the channel.

Temporal evolution of glacier hydrology and runoff

Conceptually, spatial and temporal changes in temperate glacier hydrology are thought to involve, on an annual cycle, a glacial drainage system that closes over winter and reopens during summer months.

The relatively rapid rate of ice creep in temperate glaciers, with ice at the PMP, makes it unlikely that en- and/or sub-glacial conduits or fast drainage structures remain open once empty of liquid-phase meltwater during winter. However, once closed, meltwater can become trapped through the winter in cavities at the glacier bed. At the commencement of the melt season, the volume of water stored within the glacier initially increases due to poorly interconnected drainage structures (e.g. snowpack, isolated crevasses, and flooded moulins) and the presence of a slow drainage system at the glacier bed. As surface warming continues through the spring and early summer, the surface snowpack becomes isothermal, and snowmelt floods (*spring events*) occur which flush a large volume of meltwater through the glacier drainage system, resulting in an early melt season peak in runoff volumes. This spring event may also release previously stored meltwater. As the summer progresses, ablation is heightened across the glacier surface, and increasing volumes of water are routed into the glacier interior. This increases water pressures and destabilizes slow drainage structures such that a channelized, fast flow network develops at the expense of distributed drainage (**Figures 1 and 3**). The hydraulically efficient system then probably persists for the remainder of the melt season with sufficient water flux to maintain fast drainage structures. Once ablation and meltwater volumes decline, the discrete, fast flow drainage structures begin to close through ice deformation.

[Figure 3 here]

Importantly, these temporal changes do not occur uniformly across the entire glacier bed but have a systematic spatial expression that is driven by the up-glacier expansion of the area of intense surface melting. This area corresponds to the expanding area of bare ice exposed by the melting of the (hydraulically buffered) supraglacial snowpack. The retreat of the snowline towards the ELA progressively exposes moulins and crevasses that become hydraulically active, delivering melt to en- or sub-glacial drainage flowpaths. Consequently, water pressures within and at the bed of a glacier are not spatially uniform, and the growth of a fast, discrete drainage system evolving from a slow system follows the retreat of the snowline up-glacier (**Figure 3**). The growth in extent of an efficient drainage system is also likely to reduce storage within the glacier, with a greater degree of coupling between melt processes and runoff volumes following the reduction in both snow/firn aquifer and impedance to subglacial flow. Meltwater discharge as the snowline retreats to higher elevations has been demonstrated to result in an increase in bulk meltwater runoff volumes as well as a reduction in the time lag between peak temperatures and peak runoff. As the season progresses, the diurnal runoff signal becomes increasingly peaked due to the loss of hydrological characteristics capable of damping the transport of meltwaters. Later in the melt season, with falling rates of ablation, the reduced supply of meltwater is insufficient to maintain the fast drainage structures, which begin to close and decrease in extent as they transition back to, and integrate with the distributed flow structures.

The temporal and spatial progression of a temperate glacier's conceptual hydrological system imparts systematic changes in the runoff hydrograph. Runoff may flow from a glacier throughout the year, with a *base flow* associated with slowly draining water, volumes of which only change at longer (most typically seasonal to annual) time-scales. From the onset of melt, and following any snowmelt flood events, a glacier's runoff hydrograph typically exhibits rising discharge volumes and increases in the amplitude of diurnal discharge variations as the drainage system progressively becomes more evolved and efficient (**Figure 3**). The time lag between peak melt and peak discharge at the glacier terminus is progressively reduced, and may be less than 1 hr. If melt season snowfall occurs or synoptic conditions reduce melt rates, discharge (and diurnal variations therein) are commonly diminished. As the melt season ends, the hydrograph typically shows a rapid decline in runoff volumes.

Quality of glacial runoff

For most glaciers, water emerges from the glacier terminus in a few discrete proglacial streams associated with the dominant fast subglacial flowpaths. The water quality in these streams is dependent on the flowpaths meltwater has followed; a glacier's hydrological structure is critical to the acquisition of *solutes* (ions dissolved in water) and entrainment of suspended sediment by waters emerging at the ice margin.

Solutes

The hydrochemistry of meltwaters within glacierized catchments has been studied since the 1970s. The primary hydrochemical constituents in glacial meltwaters are solutes derived from atmospheric deposition and the weathering of catchment bedrock and glacial sediments, predominantly in subglacial or ice-marginal environments. The complex system of solute acquisition by glacial meltwaters may be conveniently simplified by a two-component model whereby bulk meltwater is composed of 'quick' and 'delayed' flow. Conceptually, the former can be considered as rapid supraglacial and englacial flow, along with channelized subglacial flow, which is largely devoid of solutes, while the latter relates to slower subglacial transport pathways, enriched in solute. In practice, however, these components do not exist as discrete entities, and solutes are acquired to varying degrees by glacial meltwaters that follow composite pathways, seriously undermining the mixing model approach. However, it is well ascertained that solute acquisition is influenced by the residence time of waters at the ice-bed interface. In response to the variation in meltwater production during the summer melt seasons, typically high solute concentrations are observed at times of low discharge, with low flux and long contact times, and low concentrations during times of elevated meltwater discharge, with high flux and shorter contact times, both at the diurnal and seasonal time scales.

Precipitation in glacierized catchments is the primary source of base levels of solutes observed in runoff. Seasonal snowfall and snowpacks are, therefore, important as sources and stores of solutes within glacierized catchments. As snow begins to melt, the percolation of meltwater results in changes in that snowpack's chemical composition. Field and laboratory studies have shown that 80% of the snowpack solute load may be released within the initial ~25% of meltwater runoff. This leaching or elution of solutes may be complicated by snowpack heterogeneity. Seasonal elution has also been observed in runoff from firn. This removal of solutes from snowpack and firn, the sources of glacial ice, means the ice itself is relatively dilute. As the snowline retreats up-glacier during the ablation season, dilute ice-melt comes increasingly to dominate supraglacial runoff.

The relatively high solute concentrations measured in proglacial meltwaters must, therefore, be acquired by chemical weathering of sediments ice-marginal or subglacial environments. It is now widely recognized that hydrolysis and carbonation reactions dominate rock-water interaction and the chemical weathering within glacierized basins. Calcium ions (Ca^{2+}) are the dominant cation in glacial runoff, which reflects its relatively rapid rate of dissolution from silicate and carbonate rocks, although silica concentrations tend to be lower than in non-glacial runoff because silicate weathering is depressed in cold subglacial and ice-marginal environments. The acquisition of other base cations depends on a catchment's specific geology and geochemical susceptibility. However, observations of persistently-elevated sulphide ion (SO_4^{2-}) concentrations suggest alternative weathering reactions occur along subglacial flowpaths. Anaerobic nitrate reducing or sulphate reducing microbes and oxidising chemotrophs have been shown (or inferred) to exist in a wide range of glacial environments, and subglacial chemical weathering can be increased up to 8-fold through microbially catalysed reactions.

Sediment

The runoff from glacierised catchments characteristically exhibits high suspended sediment loads that impact upon its use and management. Consequently, similar research attention has been devoted to glaciofluvial sediment transport, which is typically confined to a limited melt season. Analyses have shown that annual specific glaciofluvial sediment yields, which routinely exceed 10^2 t km^{-2} year^{-1}, are substantially greater than global averages for other terrestrial catchments.

In view of the conceptual model of valley glacier hydrology during the ablation season (**Figure 1**), the critical distinction between the quick flowpaths of the largely debris-free supra- and en-glacial systems, with low or intermediate suspended sediment concentration (SSC), and the slower, delayed flow through the subglacial drainage system, exhibiting high SSC through entrainment of rock flour and subglacial debris, is important. The transfer of fine-grained rock flour in suspension dominates sediment evacuation from most glaciers, and particularly from temperate-based glaciers. The concentrations of sediment transported via supraglacial drainage paths, although typically < 0.5 g L^{-1}, are highly dependent on the nature and extent of debris at the glacier surface and within the ice body as well as delivery of sediment derived from extra-glacial locations, such as lateral moraines, to the supraglacial environment. Debris may be released from englacial and basal ice through the melting and enlargement of conduits by viscous heat dissipation. However, where the glacier bed is composed of unconsolidated materials, the entrainment and removal of sedimentary products will be the dominant source of suspended sediment entrained in meltwaters passing through subglacial drainage paths, with SSC commonly ranging between 1 and 10^2 g L^{-1}. Entrained sediment may itself increase SSC through mechanical erosion (*abrasion*) of bedrock or basal sediments. However, the relationship between SSC and discharge is neither linear nor stable in proglacial rivers, and sediment yields can vary markedly at diurnal, seasonal, and annual time-scales.

Changes in the rate of meltwater production at the glacier surface and the hydrological configuration within a glacier control the temporal variability in suspended sediment transport by influencing the processes of erosion, entrainment and transfer. The volume of sediment evacuated by runoff is typically viewed as a function of the availability of debris at the subglacial ice-bed interface. At a seasonal time-scale, the sediment yield relates to the evolution of the subglacial drainage system: SSC (sediment evacuation) can be high during the spring melt, when snowmelt flushes the glacier's subglacial drainage network and destabilises the slow drainage system; subsequently, up-glacier retreat of the snowline and corresponding growth of the fast subglacial drainage system may cause short-term increases in sediment delivery as the basal area accessed by subglacial meltwater increases. As the melt season progresses and diurnal melt cycles become more pronounced, sediment concentrations may be elevated as subglacial sediments are eroded by enlarging fast drainage structures. However, decreases in suspended sediment concentration may arise on a seasonal basis, arising from the exhaustion of available fines, as the subglacial drainage system spatially stabilises and then begins to close during the latter portion of the melt season.

Short-term variability in sediment transport by runoff is usually dominated by a marked diurnal cycle, typically characterised by a strongly positive relationship between discharge and SSC. However, the direct, positive association between SSC and meltwater discharge commonly breaks down. On the diurnal time-scale, *clockwise hysteresis* is frequently observed, whereby the flushing and exhaustion of sediment on the rising limb of the diurnal hydrograph occurs, and there is significantly less sediment available for transport when discharge declines. Similarly, *anticlockwise hysteresis* may occur in cases where subglacial channel water pressures reduce on the falling hydrograph limb and return flow from the distributed system elevates the SSC entrained at lowered discharges. Shorter-term sub-diurnal variations in SSC, so-called *sediment pulses*, have been related to sudden changes in drainage system connectivity over the glacier bed, to glacier dynamics that result in reconfiguration of the drainage system, and to snowmelt or rainfall events where the drainage system becomes inundated. Additional influences on glaciofluvial sediment yields from glacierized catchments may involve the reworking and erosion of typically unconsolidated materials associated with moraines, braid plains, or sandurs that typify ice-marginal environments and contemporary proglacial areas.

Summary and conclusion

The hydrology of a glacier and change therein is important for the volumes and quality of meltwater runoff. Such runoff may be of critical importance as a valuable resource but, occasionally, as an unwelcomed hazard. Our understanding of glacier hydrology has been progressively built up over recent decades; conceptually, the model of drainage pathways through mid-latitude, temperate valley glaciers remains dominant. Here, the principal controls over meltwater generation, routing and delivery have been summarized, categorizing key flowpaths as supraglacial, englacial or subglacial. Meltwater produced in the supraglacial environment drains impeded through the snowpack and/or firn, but more rapidly over bare glacier ice. While the interstitial, primary permeability of englacial environment cannot transport meltwater at velocities >1 m year^{-1}, secondary marco-scale drainage structures represent key flowpaths delivering meltwater to the glacier interior. Once meltwater reaches the bed, it follows either slow, distributed forms of drainage, or discrete, hydraulically efficient, fast channels, although these systems likely coexist and interact. Thermal and mechanical erosion can enlarge or migrate the en- and sub-glacial flowpaths while ice deformation causes their contraction. Although traditionally viewed as a top-down cascade, the supra-, en- and sub-glacial components of a glacier's hydrology are interconnected and meltwater can readily transition between these environments. Seasonally, the entire drainage system of a glacier opens, develops, becomes increasingly efficient and then begins to close once meltwater production declines. Processes of liquid meltwater storage and release occur within these three subsystems, at sub-seasonal, seasonal, and annual time-scales, and it is these processes that strongly influence the runoff volumes and water quality draining from a glacierized catchment as well as representing a strong influence on the basal motion of a glacier.

Despite the advances in research techniques that have provisioned our current knowledge of glacier hydrology, research challenges remain, particularly to forecast drainage system and hydrograph adjustments under predicted climatic forcing and environmental change scenarios. Research priorities would include: (i) refining spatial and temporal variations in the generation and routing of supraglacial meltwater and the associated role of the weathering crust; (ii) improving understanding of the architecture, function, spatial extend and temporal development of englacial drainage systems, perhaps utilizing geophysical techniques; (iii) constraining the reciprocal relationship between seasonal (and longer-term) evolution of subglacial drainage and ice mass motion. In addition to these flowpath specific research avenues, following recent awareness of microbial interaction with meltwaters, further investigation is needed into the potential role microbes may have on hydrological processes, water quality, and ecological habitat modification throughout the glacier system. Crucially, these questions remain not only for temperate valley glaciers, but also for both their non-temperate counterparts, where ice temperatures impose complexity on the manner by which meltwater is transferred through the ice body, and ice sheets, where subtle uncertainties remain over the transferability of the conceptual, temperate valley glacier hydrological model as well as the response time of hydrological structures to changes in meltwater provision.

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SEE ALSO:

Glaciers; Glacier mass balance; Glacier lake outburst floods; Mountain glaciation; Mountain hydrology; Snow; Hydroclimatology and hydrometeorology; Fluvial erosional processes and landforms; Water quality; Water resources and hydrological management

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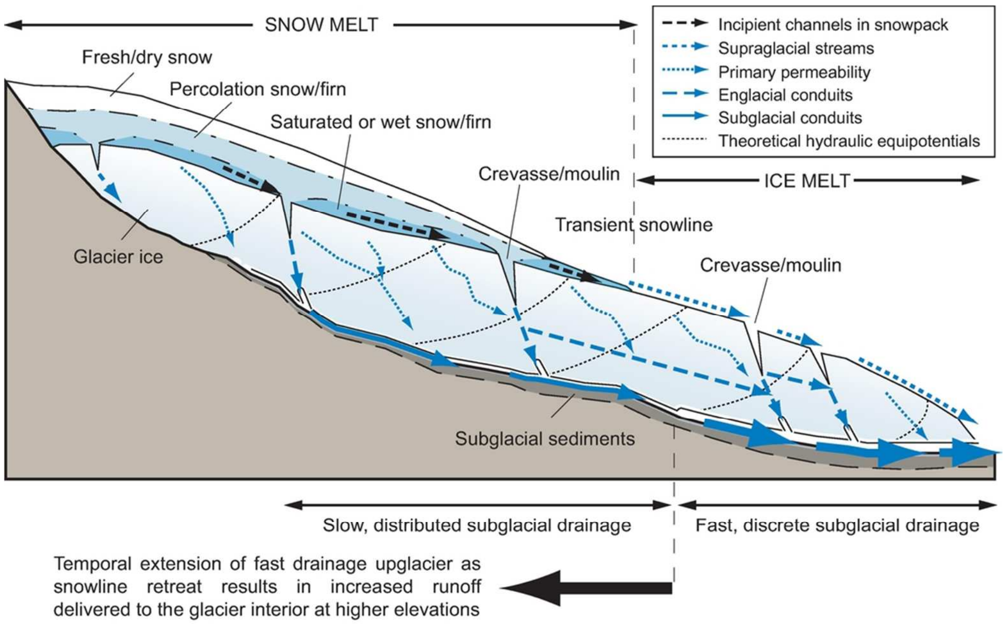
Key Words
Cryosphere, Hydrology, Topography, Geomorphology, Natural resources

Figure Captions:

Figure 1: Conceptual model of glacier hydrology. Note the drainage takes an arborescent pattern in cross-profile which is also apparent in plan view with a dendritic structure leading to the main subglacial conduit(s) or channel(s) at the glacier snout. Water flow is represented by blue arrows whose size infer discharge. [*Source*: Irvine-Fynn et al. 2011, *Reviews of Geophysics 49*(4): Figure 3].

Figure 2: Illustration of hypothetical evolution of englacial conduits from crevasses and meltwater streams. (a) Classical moulin formation from truncation of a supraglacial stream; (b) crevasse hydrofracture promotes englacial drainage path; (c) crevasse-bottom incision and glacier motion isolate an active englacial channel; (d) cut-and-closure formation of englacial channel where meandering at depth ensures englacial conduit geometry does not reflect original supraglacial stream path. Water flow is represented by blue arrows, and forces acting on ice shown by red arrows. Ice flow is left to right for (a) to (c) and from back to front in (d). [*Source*: Irvine-Fynn et al. 2011, *Reviews of Geophysics 49*(4): Figure 5].

Figure 3: Schematic of the spatial and temporal variation in meltwater travel to the proglacial environment. (a) Systematic change in tracer transit over the meltseason, showing decreasing dispersion and increasing velocity; (b) the seasonal, up-glacier extension of fast meltwater transit times mirroring the retreat of the snowline; (c) the associated seasonal runoff hydrograph. The temporal positions of idealized tracer breakthrough curves (A-D) are shown. [*Source*: combined from Richards et al., 1996, *Hydrological Processes 10*(4), Figures 2, 6 & 8, and Irvine-Fynn et al. 2011, *Reviews of Geophysics 49*(4): Figure 4].



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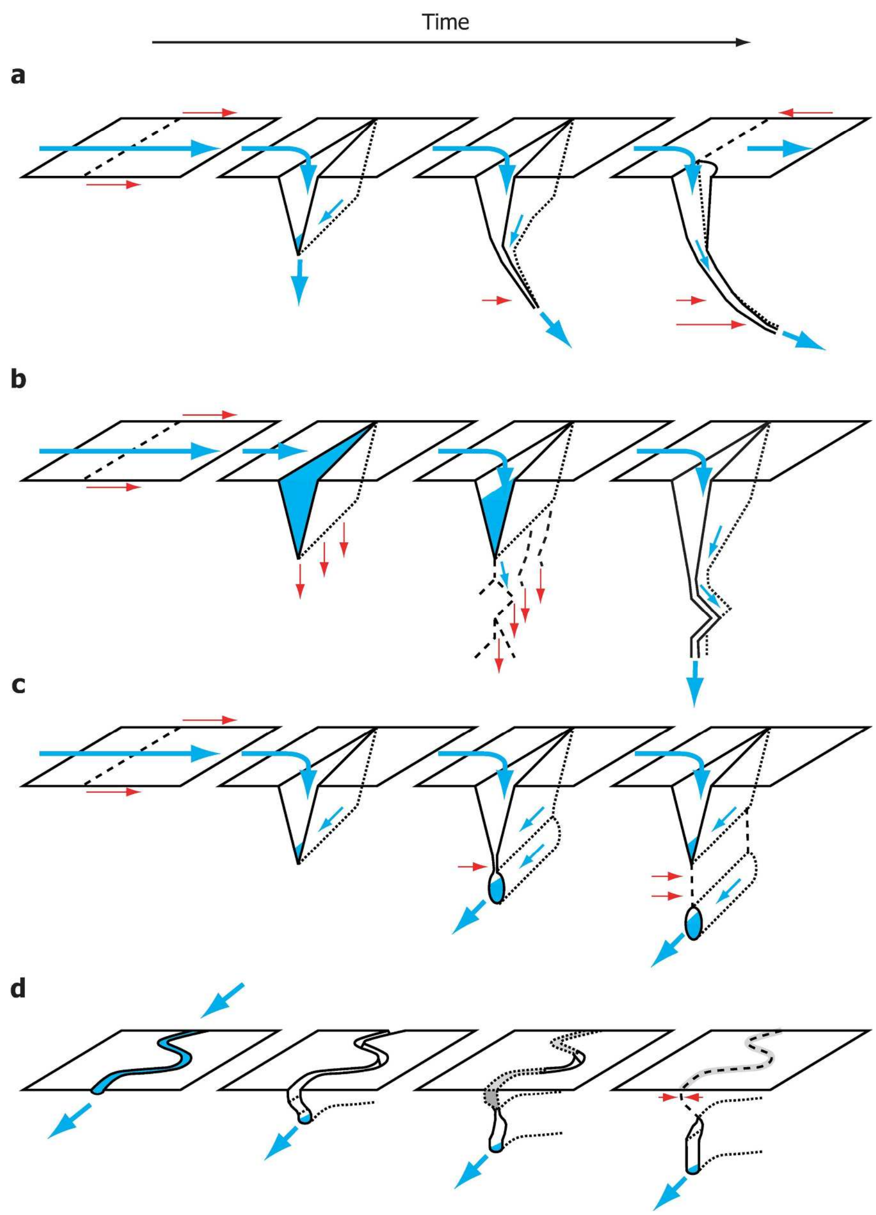

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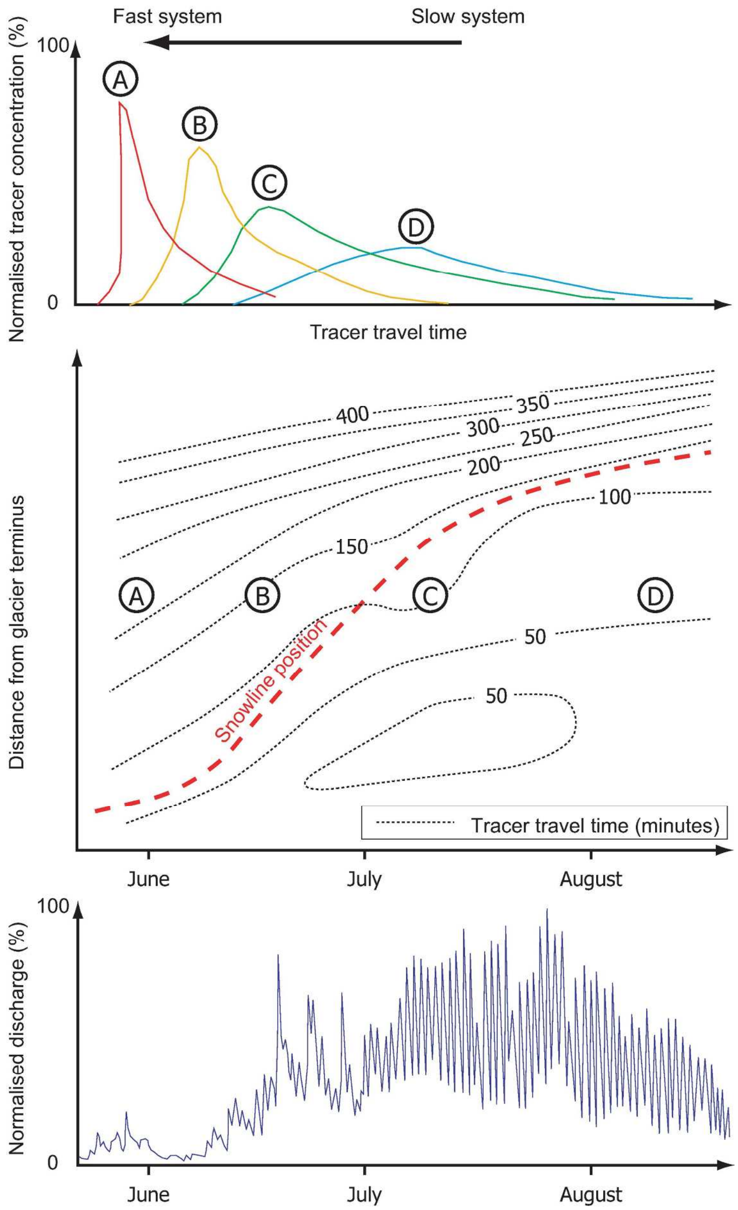

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