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Glacial Erosion Modelling

Results of a workshop held in Unterägeri, Switzerland, 29 April – 1 May 2010

December 2010

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Nationale Genossenschaft
für die Lagerung
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Telefon 056-437 11 11

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U. H. Fischer[1], W. Haeberli[2]

[1] NAGRA, Wettingen
[2] Glaciology, Geomorphodynamics & Geochronology
Geographical Institute, University of Zürich

KEYWORDS

Glacial erosion, ice-sheet modelling, erosion processes, subglacial system, abrasion, quarrying, meltwater erosion

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Summary

Deeply incised troughs and overdeepened valleys are common features in the Alps and in the Alpine foreland. These troughs and valleys as well as their sediments are of large practical and scientific interest, influencing the dynamics and hydrology of Pleistocene ice masses, presenting opportunities for glacial and climate reconstruction, and raising management issues related to aggregate, groundwater and hydrocarbon resources and radioactive waste disposal in deep geological repositories. Although the mechanisms and principles of glacial erosion are generally known, the formation of deeply incised troughs and overdeepened valleys beneath the frontal reaches of glaciers in the distal foreland of the Alps, referred to as deep glacial erosion, remains incompletely understood. In particular, the question as to when, where and how often future glaciations can lead to deep glacial erosion is of great importance for the siting and long-term safety of radioactive waste repositories in northern Switzerland.

In collaboration with the University of Zürich, the National Cooperative for the Disposal of Radioactive Waste (Nagra) organized and held a workshop in April 2010 aimed at evaluating the state-of-the-art of modelling glacial erosion as a means for developing a better understanding of the subglacial processes governing landscape evolution in the Alpine foreland of northern Switzerland in past as well as future cold environments. During the workshop a draft assessment was developed by an international group of leading experts on what quantitative information glacier erosion modelling can provide in view of safety aspects related to radioactive waste repositories in Switzerland.

During the last decade, three-dimensional models of ice flow have been developed that account for sediment erosion and transport beneath glaciers and ice sheets. The deformation and thermomechanics of glacier ice are reasonably well understood and ice cover extent and ice flow paths can be modelled with good confidence. In contrast, basal processes including glacier hydrology, ice-bed interaction, sliding, sediment transport and interactions with permafrost are comparatively poorly understood and require considerable additional model development to be usefully incorporated into comprehensive coupled models. In this respect, glacier hydrology was particularly acknowledged by the experts as being important for realistic simulations of ice dynamics and erosional processes.

There was general agreement by the experts that prognostic modelling is beyond our present capabilities and not recommended because of substantial uncertainties in the quantification of basal processes as well as the relevant climate forcing for future glacial cycles. Instead, it was suggested that diagnostic modelling has potential value for understanding processes and quantifying uncertainties and for testing sensitivities and process parameterizations.

According to the experts, modelling strategies that hold promise include ensemble modelling to explore the range of outcomes over the entire range of uncertainty in variables and extremal (end-member) modelling to bracket best and worst cases. These approaches can lead to the identification of potentially important processes and parameters and thus the component models needed (e.g. glacier hydrology, glacial sedimentary processes) for comprehensive coupled models. In deciding on an approach to the simulations, it was suggested that the relative value of ensemble runs with simpler models versus fewer runs with more complex models should be considered.

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1 Introduction

The National Cooperative for the Disposal of Radioactive Waste (Nagra) is in charge of developing deep geological repositories in Switzerland. This also includes proposing sites for such repositories and performing analyses of the long-term safety that they can provide. As part of these safety analyses the geomorphological evolution of the landscape has to be evaluated for a time horizon of one million years for the disposal of high-level waste. One of the relevant aspects concerns the effects of "deep glacial erosion". This term refers to the origin of deeply incised troughs and overdeepened valleys beneath glaciers that extended from the Alps far into the Alpine foreland and covered the midland areas of northern Switzerland (Swiss Plateau) with hundreds of meters thick ice several times during the Quaternary (Fig. 1).

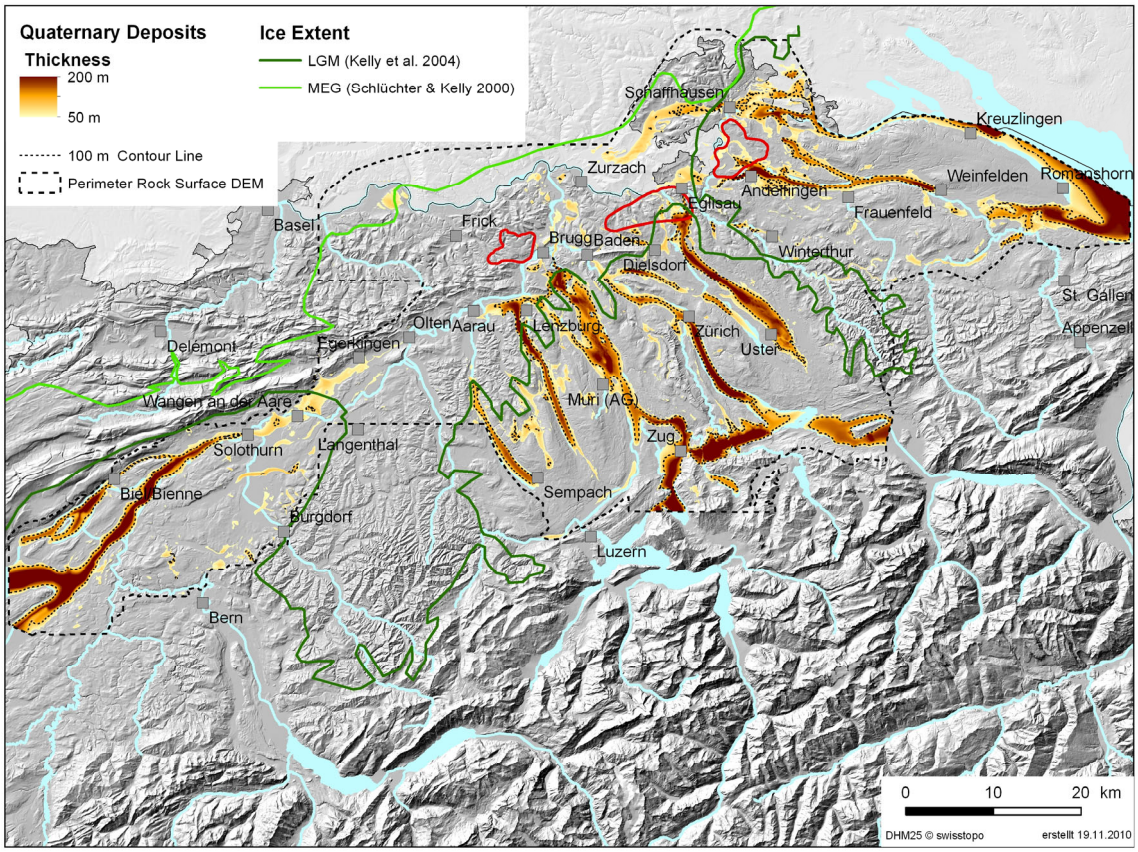

Fig. 1: Thickness of Quaternary sediments inicating a system of deep, overdeepened and buried valleys in the midland areas of northern Switzerland. Mapping is based on information from several thousand boreholes and other sources. Red polygons mark the proposed siting regions for a high-level waste repository. Also shown is the ice extent at the Last Glacial Maximum (LGM) and during the Most Extensive Glaciation (MEG).

As the climate in northern Switzerland in the time period of concern (1 Ma) is expected to continue to oscillate between glacial and interglacial periods, the question as to when, where and how often future glaciations can lead to deep glacial erosion is of great importance for the siting and long-term safety of radioactive waste repositories. Of significance in this respect is

that all of the proposed geological siting regions are located within the ice extent during the Most Extensive Glaciation (MEG), some of them even within that at the Last Glacial Maximum (LGM).

A specific workshop organized by Nagra in collaboration with the University of Zürich and held in Unterägeri, Switzerland in April 2010 was aimed at evaluating the state-of-the-art of modelling glacial erosion as a means for developing a better understanding of the subglacial processes governing landscape evolution in the Alpine foreland of northern Switzerland in past as well as future cold environments. An international group of leading experts was invited to this workshop to contribute to an assessment on what quantitative information glacial erosion modelling can provide in view of safety aspects related to radioactive waste repositories in Switzerland.

Prior to the workshop all experts were supplied with the following reports and papers providing available information relevant to the focus of the workshop:

Fischer, U.H. (2009): Glacial erosion: a review of its modelling. Nagra Arbeitsbericht NAB 09-23.

Haeberli, W. (2010): Glaciological conditions in northern Switzerland during recent Ice Ages. Nagra Arbeitsbericht NAB 10-18.

Herman, F. & Braun, J. (2008): Evolution of the glacial landscape of the Southern Alps of New Zealand: insights from a glacial erosion model. Journal of Geophysical Research, 113, F02009, doi:10.1029/2007JF000807.

Hildes, D.H.D., Clarke, G.K.C., Flowers, G.E. & Marshall, S.J. (2004): Subglacial erosion and englacial sediment transport modelled for North American ice sheets. Quaternary Science Reviews, 23, 409-430.

Jamieson, S.S.R., Hulton, N.R.J. & Hagdorn, M. (2008): Modelling landscape evolution under ice sheets. Geomorphology, 97, 91-108.

Jordan, P. (2008): Designing the DEM of the base of the Swiss Plateau Quaternary sediments. Proceedings of the 6th International Cartographic Association (ICA) Mountain Cartography Workshop "Mountain Mapping and Visualisation", 107-113.

Pollard, D. & DeConto, R.M. (2003): Antarctic ice and sediment flux in the Oligocene simulated by a climate-ice sheet-sediment model. Palaeogeography, Palaeoclimatology, Palaeoecology, 198, 53-67.

Preusser, F., Reitner, J.M. & Schlüchter, C. (2010): Distribution, geometry, age and origin of overdeepened valleys and basins in the Alps and their foreland. Swiss Journal of Geosciences, 103(3), 407-426, doi:10.1007/s00015-010-044-y.

This report is a synthesis of the main results of the workshop. Chapter 2 lists all the participants of the workshop. In Chapter 3, the organisation of the workshop is described and an overview of the program is given. Chapter 4 presents the key questions that were used to guide the discussions as well as the corresponding answers that were developed together with the experts. Finally, abstracts of all the presentations given by the experts are compiled in the Appendix.

2 Participants

Jens K. Becker	National Cooperative for the Disposal of Radioactive Waste (Nagra) Hardstrasse 73 5430 Wettingen Switzerland jens.becker@nagra.ch
Simon H. Brocklehurst	School of Earth, Atmospheric and Environmental Sciences University of Manchester Manchester M13 9PL United Kingdom shb@manchester.ac.uk
Garry K.C. Clarke	Department of Earth and Ocean Sciences University of British Columbia 6339 Stores Road Vancouver, BC V6T 1Z4 Canada clarke@eos.ubc.ca
Denis O.H. Cohen	Institut für Terrestrische Oekosysteme ETH Zürich Universitätstrasse 16 8092 Zürich dcohen@iastate.edu
David L. Egholm	Department of Earth Sciences University of Aarhus Høegh-Guldbergs Gade 2 8000 Aarhus C Denmark david@geo.au.dk
Todd A. Ehlers	Institut für Geowissenschaften Universität Tübingen Wilhelmstrasse 56 72074 Tübingen Germany todd.ehlers@uni-tuebingen.de
Derek Fabel	Department of Geographical and Earth Sciences University of Glasgow Gregory Building University Avenue Glasgow G12 8QQ United Kingdom derek.fabel@ges.gla.ac.uk
Urs H. Fischer	National Cooperative for the Disposal of Radioactive Waste (Nagra) Hardstrasse 73 5430 Wettingen Switzerland urs.fischer@nagra.ch

Gwenn E. Flowers
Department of Earth Sciences
Simon Fraser University
8888 University Drive
Burnaby, BC V5A 1S6
Canada
gflowers@sfu.ca

Erik Frank
Eidgenössisches Nuklearsicherheitsinspektorat (ENSI)
Industriestrasse 19
5200 Brugg
Switzerland
Erik.Frank@ensi.ch

Brent M. Goehring
Lamont-Doherty Earth Observatory
Columbia University
61 Route 9w
PO Box 1000
Palisades, NY 10964
U.S.A.
goehring@ldeo.columbia.edu

Wilfried Haeberli
Geographisches Institut
Universität Zürich
Winterthurerstrasse 190
8057 Zürich
Schweiz
wilfried.haeberli@geo.uzh.ch

Bernard Hallet
Quaternary Research Center
Department of Earth and Space Sciences
University of Washington
Box 351360
Seattle, WA 98195-1360
U.S.A.
hallet@u.washington.edu

Frédéric Herman
Geologisches Institut
ETH Zürich
Sonneggstrasse 5
8092 Zürich
Switzerland
frederic@erdw.ethz.ch

Nicholas R.J. Hulton
School of Geosciences
University of Edinburgh
Grant Institute
The King's Buildings
West Mains Road
Edinburgh EH9 3JW
United Kingdom
Nick.Hulton@ed.ac.uk

Neal R. Iverson
Department of Geological and Atmospheric Sciences
Iowa State University
253 Science I
Ames, IA 50011
U.S.A.
niverson@iastate.edu

Stewart S.R. Jamieson
Department of Geography
Durham University
Science Laboratories
South Road
Durham DH1 3LE
United Kingdom
Stewart.Jamieson@durham.ac.uk

Michele N. Koppes
Department of Geography
University of British Columbia
1984 West Mall
Vancouver, BC V6T 1Z2
Canada
koppes@geog.ubc.ca

Joachim Kuhlemann
Eidgenössisches Nuklearsicherheitsinspektorat (ENSI)
Industriestrasse 19
5200 Brugg
Switzerland
Joachim.Kuhlemann@ensi.ch

Kelly R. MacGregor
Geology Department
Macalester College
1600 Grand Avenue
St. Paul, MN 55105
U.S.A.
macgregor@macalester.edu

David Pollard
Earth and Environmental Systems Institute
Pennsylvania State University
College of Earth and Mineral Sciences
2217 Earth-Engineering Sciences Building
University Park, PA 16802-9297
U.S.A.
pollard@essc.psu.edu

Frank Preusser
Institut für Geologie
Universität Bern
Baltzerstrasse 1+3
3012 Bern
Switzerland
preusser@geo.unibe.ch

Jürgen M. Reitner
Geologische Bundesanstalt (Geological Survey of Austria)
Abteilung Sedimentgeologie
Neulinggasse 38
1030 Wien
Austria
juergen.reitner@geologie.ac.at

Christian Schlüchter
Institut für Geologie
Universität Bern
Baltzerstrasse 1+3
CH-3012 Bern
Switzerland
schluechter@geo.unibe.ch

Michael Schnellmann
National Cooperative for the Disposal of Radioactive Waste (Nagra)
Hardstrasse 73
5430 Wettingen
Switzerland
michael.schnellmann@nagra.ch

Dorothea Stumm
National Cooperative for the Disposal of Radioactive Waste (Nagra)
Hardstrasse 73
5430 Wettingen
Switzerland
dorothea.stumm@nagra.ch

Darrel A. Swift
Department of Geography
University of Sheffield
Winter Street
Sheffield S10 2TN
United Kingdom
D.A.Swift@Sheffield.ac.uk

Mauro A. Werder
Department of Earth Sciences
Simon Fraser University
8888 University Drive
Burnaby, BC V5A 1S6
Canada
mawerder@sfu.ca

Brian J. Yanites
Department of Geological Sciences
University of Michigan
C.C. Little Building
1100N. University Avenue
Ann Arbor, MI 48109
U.S.A.
yanites@umich.edu

Piet Zuidema
National Cooperative for the Disposal of Radioactive Waste (Nagra)
Hardstrasse 73
5430 Wettingen
Switzerland
piet.zuidema@nagra.ch

3 Program

The workshop was organised to have an introductory part on the morning of 29 April, describing the general problems and goals, the development in time of Alpine foreland glaciation and paleoclimatic/paleoglaciological conditions during maximum ice extent on the Swiss Plateau. This was followed by an afternoon session with presentations and discussions about observations and measurements related to glacial erosion. Following a short excursion on the morning of 30 April, the afternoon was devoted to a session with presentations and discussions about possibilities and challenges of glacial erosion modelling in view of future ice ages. The presentations were short introductions with the idea to acquaint all participants with the current state of research in the different areas of expertise (see Appendix for abstracts). The discussions were structured in a way that led to answers to key questions that had been formulated prior to the workshop (see Chapter 4). The morning of 1 May was set aside for the final discussion and integration of the main outcomes of the workshop.

28 April:

Evening	Arrival of Participants (SeminarHotel am Ägerisee)

29 April:

Time / Session	**Speaker**	**Title of Presentation**
8:30	P. Zuidema	Welcome address
	P. Zuidema	Radioactive waste disposal in Switzerland
9:15 Introduction & Context	U. Fischer	Introduction to the workshop
	M. Schnellmann, A. Gautschi	Background and past research activities with respect to deep glacial erosion in north-eastern Switzerland
	F. Preusser, C. Schlüchter, J. Reitner	Quaternary glaciation history of the Alps
	W. Haeberli	Glaciological conditions in northern Switzerland during recent ice ages
	B. Hallet	Constraining the plausible amount of future glacial erosion
12:00	*Lunch*	

29 April:

Time / Session	**Speaker**		**Title of Presentation**
13:30 Session 1: Measurements & Observations	B. Hallet		Introduction, Questions
	D. Fabel	10’	Quantifying glacial bedrock erosion with cosmogenic nuclides
	B. Goehring	10’	A new method to estimate glacial erosion rates using in situ 14C and 10Be: an example from the Rhone Glacier
	M. Koppes	10’	Empirical measurements of basin-averaged glacier erosion rates beneath calving glaciers
	J. Braun (presented by F. Herman)	10’	Use of low-temperature thermo-chronology to quantify glacial erosion
	S. Brocklehurst	10’	Insights into glacial landscape evolution from digital topographic analyses
	D. Cohen	10’	Field evidence for water pressure transients increasing rates of quarrying
	W. Haeberli	10’	Rock and sediment beds of mountain glaciers – an alternative (quantitative) approach
	D. Swift	10’	Sediment evacuation by subglacial meltwater and the implications for glacial erosion
			Discussion
18:00	*Dinner*		

30 April:

Time / Session	**Speaker**		**Title of Presentation**
8:30	C. Schlüchter, W. Haeberli		Excursion
11:30	*Lunch*		
13:00	G. Clarke		Introduction, Questions
Session 2: Modelling	G. Clarke	10’	Computer modelling of mountain glaciation
	D. Egholm	10’	Modelling glacial erosion with a higher order ice-sheet model
	F. Herman	10’	Effects of glacier hydrology and sediment transport on erosion patterns
	T. Ehlers	10’	Influence of various periglacial and glacial processes in alpine glacial erosion model
	S. Jamieson	10’	Modelling long-term glacial landscape evolution: feedbacks between ice, erosion and topography
	N. Hulton	10’	Modelling glacial erosion: heat, friction and water
	K. MacGregor	10’	Simple modeling of quarrying and abrasion along the glacier longitudinal profile
	N. Iverson	10’	A model of quarrying based on adhesive wear and Weibull rock fracture
	G. Flowers	10’	Progress in modelling glacier and ice-sheet hydrology
	M. Werder	10’	Water flow speed in the channelised drainage system
	D. Pollard	10’	Coupling deformable sediment models with ice sheet models
			Discussion
18:30	*Dinner (Gasthof Gubel)*		

1 May:

Time / Session	**Speaker**	**Title of Presentation**
8:30 Integration & Discussion	B. Hallet, G. Clarke	Introduction, Summary
		Approaches to evaluating erosion in the Alpine foreland – editing of the expert assessment
12:00	*Lunch*	
Afternoon	Departure of Participants	

4 Expert answers to key questions

The following questions were used to guide the discussions during Session 1 (Measurements & Observations) on 29 April and Session 2 (Modelling) on 30 April. The answers were compiled during the final discussion and integration on 1 May and reflect the main outcomes of the workshop.

4.1 Session 1: Measurements & Observations – Processes and rates of glacial erosion

(1.1) Both, erosion and sedimentation by glaciers are possible: What factors influence corresponding processes and under what conditions are rock and sediment beds likely to develop?

Sub- and proglacial meltwater is essential for sediment evacuation, cliffs adjacent to the ice surface and subglacial uptake provide for debris input. Integrated consideration of the entire system is necessary, including aspects of the sediment balance and history (fluctuations, glacial/periglacial/fluvial sequences). The primary question is whether sediment is retained within glacial transport pathways or evacuated beyond the catchment area by glacio-fluvial processes. The latter will be favourable for bedrock erosion, as it will enhance ice-bedrock interaction. If glacio-fluvial evacuation is not efficient, sediment may accumulate to thicknesses where it starts to limit glacial erosion of bedrock. There is widespread evidence for the dominance of glacio-fluvial processes in basal sediment evacuation. The distribution of rock and sedimentary beds is likely to be controlled broadly by the efficiency of glacio-fluvial sediment evacuation in relation to sediment production by bedrock erosion.

Favourable/*unfavourable* factors for erosion into bedrock are

- temperate/*cold* basal ice
- large/*small* glacier area
- high/*low* surface/bed slope
- small/*large* cliffs

Surface slope is important as glacial erosion is coupled to ice-velocity which in turn is coupled to surface slope. Further effects are due to changes in lithology along transport paths and to the occurrence of pre-existing (undisturbed or reworked) sediments, which can affect bedrock erosion and may confound estimates of erosion rates that are based on sediment yields. Key questions concern details of long-distance transportation: Where are the sediments generated and where are they deposited?

(1.2) Overall denudation of glacier catchment-areas is caused by glacial erosion and periglacial cliff recession: What are the relative contributions of the two?

The relative contributions by glacial and periglacial (sub-aerial) processes depend on the ratio of total glacier area to the extent of exposed cliffs in the catchment area. Periglacial cliff recession is recognizable from angular components in sediment deposits and supra- and englacial fractions of debris content in glacier ice, which typically do not amount to more than 10 % per unit volume of ice. Such products can represent an important contribution to overall catchment-area denudation, particularly during phases of deglaciation, as the area of exposed rock walls

increases and oversteepened slopes become debuttressed as ice thins and retreats. In cases of relatively small glaciers in large-relief catchments, periglacial erosion may even predominate over spatially averaged glacial erosion. However, in icefields such as in Patagonia today, or during glacial maxima in the Alps, when very little of the ridges rise above the ice surface, periglacial erosion and storage may be a negligible component of overall denudation.

Existing data on catchment-area denudation rates should be systematically analyzed with respect to such effects and with respect to time-transgressive changes in ice thickness and relief production to better understand the processes involved and to enable more reliable extrapolations to unmeasured regions. Some means of doing so include quantifying landslide volumes on shrinking glaciers and comparing those to outlet sediment yields, quantifying headwall retreat rates using cosmogenic exposure dating, and/or incorporating frost-cracking rates and rock-wall areas into models for estimating periglacial contributions and the potential flux of rock tools from the headwall through the bergschrund to the bed.

(1.3) What factors and processes control the overall sediment balance (input-transfer–output)? Where is subglacial sediment eroded, where is it transported to and where is it deposited?

Uptake at the bed (abrasion, plucking), transport (by ice or water flow) and (temporary) deposition of debris can critically influence basal friction, sliding and abrasion by ice but remain difficult to estimate accurately. Sediment balance should be considered in connection with basal debris concentration, subglacial hydrology and basal sediment mobilisation. Control on basal sliding, headwall processes and debris incorporation in ice are among the most prominent open questions.

The spatial distribution of erosion of bedrock will be strongly influenced by the efficiency of glacio-fluvial evacuation of sediment. The importance of hydrology in sliding rates, quarrying and the efficiency of sediment evacuation means that the greatest erosion rates should occur downglacier of the equilibrium line altitude (ELA) position. The efficiency of glacio-fluvial evacuation of sediment will also control the partitioning of sediment between glacial and fluvial sediment transport pathways, and the presumption is that glacio-fluvial evacuation is dominant, such that most sediment will be evacuated from the catchment area. Only a small proportion of sediment is likely to remain in glacial transport to be deposited by glacial processes (e.g. in tills/moraines). Better knowledge is needed about the importance of the processes over long time scales of interest (in the case of repositories ~ 10^6 years) with complex effects from changing glacier geometries and climatic conditions.

(1.4) Glacial erosion comprises large-area denudation by subglacial abrasion versus concentrated down-cutting by subglacial water channels. What are the relative contributions of the two processes and what factors are important for the spatial and temporal distribution of corresponding landforms? Where in the glacial system and at what time in a glacial cycle are they most relevant?

Subglacial abrasion and quarrying are pervasive in bedrock areas covered by warm-based glaciers, and may be largely responsible for the overall eroded volumes, especially where the ratio of total glacier area to the extent of exposed cliffs is large. Concerning maximum possible erosion depth, locally concentrated erosion by subglacial meltwater may have to be taken into account. Evidence in the field covers a wide range of landforms from deep/narrow (formerly subglacial) gorges, to deep V-shaped, glacially polished fjords and troughs, to wide valleys with broad cross-sections. Ice flow in narrow valleys/channels probably favours strong and locally

concentrated subglacial meltwater erosion. Locally concentrated erosion by subglacial meltwater has indeed created deep/narrow gorges in selected places. However, these locations appear to have specific characteristics that are not very likely to be re-created in the lowland areas under consideration. These include, in general, steep ice and bed surfaces that favour locally focused erosion because of the very high subglacial hydraulic potential gradient that drives the water. Topographic, paleoglaciological and lithological/structural settings of known deep/narrow (formerly subglacial) gorges need systematic investigation. Evidence for the rarity or absence of vigorous localized erosion by subglacial meltwater include bedrock riegels without traces of strong fluvial erosion, and extensive areas of glacially scoured bedrock with little or no trace of incised channels. Ice flow in narrow valleys/channels probably favours strong and locally concentrated, rapid subglacial meltwater erosion.

In the past, subglacial abrasion and quarrying have taken place over large linear areas, creating pronounced overdeepenings in the foreland. Multiple glaciations have re-occupied the same troughs and may continue to do so in the future. Concerning integrated effects over large areas, the contribution of glacial erosion likely dominates that of focused meltwater incision. The latter may, however, play an important role with respect to the greatest depths reached. Selective linear glacial erosion appears to dominate just inboard of the LGM ice margin in northern Switzerland, where warm-based conditions are expected.

(1.5) How does the resistance to erosion of different lithologies (consolidated rock, lake deposits, unconsolidated Quaternary deposits) affect erosion rates?

Observations show a dependence of the position and shape of deep glacial valleys on lithology, extent of fracturing, and fault zones. In contrast, most of known, very narrow gorges are located within competent rocks (mainly limestones). Hardness influences abrasion. Rock strength, which is strongly influenced by pre-existing fractures and flaws are important for quarrying. In highly permeable rock (e.g. karstified limestone or other lithologies that are highly fractured) it is difficult to build up the high basal water pressure that is favourable for active quarrying, and the energetic water flow necessary for sediment evacuation. Dissolution may significantly add to the mechanical erosion by water in limestone. Tills and gravel can prevent bedrock erosion until the sediments are evacuated. Loose fine-grained material can be mobilised and evacuated at extreme rates, most probably by subglacial rivers but perhaps also by rapid deformation of soft sediment beds.

Quantitative site-specific data on rates of bedrock erosion and sediment entrainment are sparse, at best. One of the priorities for future research is to examine how ice and water interact with sediments under a glacier.

(1.6) What are typical rates for different erosional processes? How important is the distinction between continuous versus episodic erosion events?

Average catchment-area denudation rates over long time periods depend on the tectonic situation, climatic conditions, etc. Long-term rates are therefore best locally estimated. They can be orders of magnitude lower than decadal rates in regions where tectonic activity is minimal or crustal convergence does not sustain high rock uplift rates. Short-term rates (1-50 years) can be very fast and are important for process understanding. Typical rates of basin-wide erosion have been empirically related to depth-averaged ice velocity, with recent rates measured during post-Little Ice Age glacier retreat exceeding longer-term erosion rates (over one glacial advance-retreat cycle) by up to a factor of 5. For fast moving, temperate calving glaciers, basin-wide

erosion rates may exceed 1 cm/yr, with up to 10 cm/yr of erosion during annual peaks in return. For small, alpine catchments and polar regimes, basin-wide erosion rates drop to < 0.5 mm/yr.

Abrasion can be very slow, but lateral variations in abrasion rates can be large where there are large lateral variations in basal ice speeds or in the concentration and thickness of basal debris. Using exposure dating with cosmogenic nuclides as well as independent geologic observations, abrasion is typically much slower than quarrying under the same glacial conditions. Abrasion rates generally increase with the inferred sliding rate. At Rhone glacier, for instance, they average 0.5 mm/yr since the Late Glacial Maximum, reaching a maximum of 2 mm/yr in the center of the glacial trough but only 0.2 mm/yr at the valley sides. Quarrying rates appear to be much greater, often resetting the ages from exposure dating, i.e. > 2.5 m of erosion.

When comparing erosion rates measured over various times scales, such as using annual sediment yields, radiocarbon, exposure ages and low-temperature thermochronometry, glacial erosion rates tend to decrease by 1-2 orders of magnitude from decadal to longer time scales (1 Ma) in regions where tectonic activity is minimal or crustal convergence does not sustain high rock uplift rates. A similar decrease is also observed in areas of active uplift, but it is much less. This suggests that (i) a distinction between stochastic and continuous events needs to be made, and (ii) current measures of glacial erosion are exceptionally high, representing an unusual period of dynamic response.

(1.7) In what situations (topography, lithology, ice condition, climate) do overdeepenings typically occur? To which erosional processes can they be attributed?

The largest overdeepenings known are in fjords eroded by temperate or at least warm-based glaciers that end in marine waters and which contain abundant meltwater that has access to the glacier bed and efficiently evacuates erosional products. Ice confluences, the base of cirques and of headwalls are characteristic topographic features where overdeepenings can be found on land. Soft lithologies, strongly fractured rock and steady glacier positions further help. Overdeepenings close to the margin would require strong erosion underneath the glacier front and margins to remain stable for a long time period, whereas excavation of overdeepenings upglacier of the margin could be quite independent of glacial margin stability.

4.2 Session 2: Modelling – Ice flow and basal processes

(2.1) Are ice-flow models based on the shallow ice approximations sufficient for the simulations of glaciations on the complex topography of the Alps or are higher-order models required?

Stress gradients are important where ice flow velocities show a high degree of variation, independent of whether the variation is due to topography, hydrology or the presence of subglacial sediment. The topographic gradients of the Alps make it necessary to reflect on the validity of the shallow ice approximation (SIA). Theoretically, the SIA is only valid for very small bed slopes (< 0.01), while higher-order models are accurate for higher slopes (< 1 for the second-order approximation) and full-Stokes methods will give an even more accurate representation of the glacial dynamics. SIA ice sheet models, e.g., consistently overestimate the extent of ice in areas with high topography. The higher-order effects that are neglected by the SIA impede the flow through narrow valleys and perturb the basal shear stress, with consequences for predicted glacial erosion patterns. Further, higher-order physics is important to incorporate the thermodynamics appropriately. For example, problems associated with the

strain heating term in SIA models are resolved if a higher-order model is used. In the alpine foreland, bed slopes are much smaller and the SIA should intuitively be more valid, although the flow of ice in the foreland may be influenced by higher stress gradients from the ice in the alpine topography.

Higher-order ice sheet models are becoming standard in the modelling community and many versions are freely available. This could turn into a credibility issue and it therefore would seem safest to use higher-order models as the opportunity is there and will increase in the near future. However, if computing is limited and full-stress runs are computer-intensive ensemble modelling using SIA models is likely to deliver a greater benefit than significantly fewer runs with higher-order stress treatment. In this respect, one must be clear on the scales (space and time) of interest and consider the limitations of applying the SIA relative to other uncertainties in models such as the boundary conditions (subglacial hydrology, basal temperature regime, mass balance, thermal meltwater effects, debris cover).

(2.2) What processes must be included in models of glacial erosion? How can the essential processes be best included? What is the appropriate approach to simulations?

Hydrology is generally acknowledged as being important for realistic simulations of ice dynamics and erosional processes, which are likely important for Nagra's application. For example, bedrock erosion by pressurized subglacial water is probably essential for excavating deep gorges such as can be found in Switzerland today (e.g. Aareschlucht). However, it is unclear how the deep gorges form (e.g. whether subglacially or proglacially) and to what extent they are structurally determined vs. determined by transient conditions associated with the glacier or the underlying landscape. Base-level lowering may be relevant. Regardless, these features are much smaller than can be resolved in distributed models, and could perhaps be treated off-line by constructing worst case scenarios (e.g. for slot depth) and looking at modern analogues.

Permafrost may be important in future glaciations as it has been a feature of the glacierized landscape in the past, and may produce unwelcome surprises in its ability to dam water that could be released catastrophically or activate deep thrusting. Permafrost thus may play an important role in basal sediment entrainment which would be relevant to Nagra's interest, but the relevant processes are not included in current ice-flow models.

Climate interacts with glaciers primarily through the surface mass balance, which acts as a first-order control on ice-mass geometry and dynamics. Some effort should therefore be devoted to producing reliable mass-balance models. Reanalysis data and downscaling could be employed in an effort to provide relevant climate drivers at the appropriate scale for mass balance models, depending on the simulation approach taken.

Models that attempt to couple processes that operate on disparate timescales (e.g. hydrology) should make judicious use of asynchronous coupling with ice dynamics, and might benefit from statistical characterizations of key processes (e.g. temporal variability in basal water pressures on timescales too short to explicitly model) and parameterizations based on off-line models of subgrid processes (e.g. calculating the fraction of the year that the drainage system is conduit-dominated). In this respect, a site-specific approach may help focus the effort and guide the selection of essential processes and parameterizations. In deciding on an approach to the simulations, one might consider the relative value of ensemble runs with simpler models vs. fewer runs with more complex models.

In constructing a strategy for driving the models, one should consider whether uniformitarianism can be applied in envisioning future climate (e.g. continuation of the 100 ka ice-age cycle for the time period of interest). Trying to project future climate on the 1 Ma timescale seems misguided, and thus one should consider alternatives such as constructing end-member models (of either climates or glacial conditions) and looking for modern glacial analogues to these, examining paleo-climate proxies such as the pollen record for temperature and precipitation to guide the choice of forcings, or take a stochastic/Bayesian approach in which probability density functions would replace individual realizations of climate.

(2.3) What is the likelihood of new valleys forming nearby existing troughs under conditions of selective linear erosion of polythermal glaciers? Under what circumstances does the advancing ice at the onset of glaciations exploit an existing valley network or carve out new valleys? Will lateral ridges remain essentially intact during future glacial advances?

In general, a glacier will easily erode if existing conditions will allow it: Deeps get deeper because (i) ice and water are steered into favourably oriented preexisting troughs and away from hills and ridges and (ii) for a polythermal glacier, more effective erosion occurs under thick ice (all else being equal) because it is more likely to be at the melting point. Furthermore, erosion and trough initiation may be induced by heat advection from upglacier which propagates down the valley.

The likelihood of new valleys forming nearby existing troughs is influenced by the future climate evolution and associated amplitude of future glaciations. A key aspect in this respect is a perturbation resulting from spatial shifts in accumulation centres. Such shifts could occur because of both shifting ice divides (glacier capture, equivalent to river capture) and shifting spatial patterns of mass balance due to changes in climate.

On a more local scale, additional factors, such as the relief of the existing landscape, the lithology of the substrate and the orientation of existing valleys relative to the ice flow direction, may influence the tendency for new valley formation.

(2.4) Can the deeply incised troughs and overdeepened valleys in LGM ice-marginal zones of northern Switzerland essentially be cut in a single glacial cycle or does their formation require further deepening of existing valleys during subsequent glaciations?

There is good evidence to suggest that tunnel valleys e.g. in northern Germany or Denmark were cut in a single glaciation. However, it is not clear whether tunnel valleys are an appropriate analogue to the troughs and overdeepened valleys in northern Switzerland and whether they were formed by the same processes. Nevertheless, nature provides considerable guidance in defining maximum rates of bedrock incision in large valleys. The maximum depth of fjords, which is > 1 km, as erosion of bedrock below LGM sea level is most likely to reflect glacial erosion during the Quaternary (2-3 Ma period). This points to an upper limit for the erosion of bedrock in tectonically inactive areas such as stable continental margins, like Labrador, Greenland and Norway, where the long-term crustal uplift and erosion of the uplands are very low (0.3-0.5 mm a^{-1}). The depth of fjords in actively uplifting areas is confounded by the regional uplift and erosion.

Under the right conditions, subsequent reoccupation of valleys by glaciers can lead to further erosion, however at a slower rate. The key question is what can reduce or shut off erosion as

subsequent reoccupation occurs. There is a strong suggestion that erosion rates can decrease after initial formation but it is unclear exactly what limits the overall depth to which troughs can be over-deepened. Bedrock incision requires the ice to be in forceful contact with the bed (as opposed to floating) and for the sediment to be removed from the overdeepening. For temperate glaciers, the latter is likely to be dominated by subglacial fluvial erosion and entrainment of sediment, requiring that conditions sustain vigorous subglacial flow of water. Both this water flow and the firm contact at the ice-bed interface (non-zero effective pressure) are largely determined by the geometry of the glacier and bed.

(2.5) How can the origin of deeply incised troughs and overdeepened valleys in the LGM ice-marginal zones of northern Switzerland be explained? What processes are important for deep glacial erosion beneath frontal reaches of glaciers far away from the Alps?

For cold and polythermal glaciers, the major potential mechanism that can lead to deep glacial erosion beneath their frontal reaches is considered to be the focusing of basal heat production and downglacier heat transport into these zones. Given constant basal velocity, the former is caused at the point where the product of surface slope and ice thickness is maximised. This in turn produces favourable conditions for basal sliding and strong erosion with subsequent additional feedback. Fast flow focused in topographic constrictions upglacier is likely to persist as a focused feature into the foreland area.

In contrast, for temperate glaciers, volumes of meltwater are likely to increase towards the terminus providing efficient mechanisms for subglacial fluvial erosion and sediment evacuation. However, if glaciers are frozen at their tongues, increasing volumes of meltwater towards the terminus are questionable as a cause for increased erosion.

Troughs may have also primarily formed prior to the LGM in warmer conditions, including during previous glaciations, so that most erosion occurred when glaciers were wet-based. The coincidence of the downglacier end of the troughs with the molasse/carbonate limestone may also be significant.

(2.6) How can the filling of glacially eroded valleys by glacio-fluvial sediments be incorporated into models for simulating effects from repeated glacial cycles?

End-member scenarios could be modelled, e.g. complete filling of the overdeepenings. However, whether overdeepenings become completely refilled or not depends on whether deglaciations occur through in situ wastage or active retreat. Actively retreating glaciers continue to erode bedrock and are thus likely to produce larger sediment volumes than glaciers that down waste.

For accurate modelling of the glacial and hydrological processes that transport and deposit sediment and that control sediment size and the sedimentary architecture of the deposit, it is important (i) to improve our understanding and ability to model how ice and water interact with the subglacial sediment and (ii) to develop models of subglacial hydrology that include sediment entrainment and transport.

(2.7) What situations for the formation of overdeepenings (e.g. confluence of glaciers, change in bedrock lithology, down-glacier of bedrock protuberance) are currently being reproduced by modelling?

Overdeepenings are easily produced at the confluence of glaciers in the interior of mountain ranges as increases in ice flux locally cause increased sliding and therefore erosion. Models with a thermomechanical treatment of ice flow can adequately simulate melting conditions and generation of meltwater at the bed, which increases sliding and therefore erosion, allowing an overdeepening to form. Locations of preexisting overdeepenings typically continue to erode in most 3D models, and stop only when negative feedbacks (see 2.8 below) are imposed. While not modeled specifically, it is agreed it would be relatively easy to produce overdeepenings by simply changing the relative erodibility of different parts of valleys/landscapes. In this respect, exisiting models can be powerful tools for quantifying the relative erodibility of different lithologies. Further, there is a general feeling that bed heterogeneities anywhere in a valley or even lowland environment could potentially trigger the initiation of an overdeepening.

(2.8) Are there positive and/or negative feedback mechanisms related to the formation of overdeepenings?

A resounding yes to the presence of feedbacks. Overdeepenings would appear to be the default state of the glacier bed. The examples in the Alpine foreland of northern Switzerland are not in confluence situations and, hence, there must be strong factors favouring the formation of overdeepenings.

The main positive feedbacks are topographic steering of ice into an incipient overdeepening and crevassing at the glacier surface at the upglacier end of the overdeepening that will direct surface meltwater to the glacier bed in these locations. The latter is important for sliding, quarrying and sediment evacuation. Negative feedbacks include longitudinal stresses that would prevent acceleration of ice into the overdeepening and supercooling. It should be noted that supercooling will not necessarily limit the depth of overdeepening, but it will control the angle of the reverse slope. It may also discourage efficient meltwater flow and sediment evacuation along the bed of the overdeepening. The latter effect would enable build-up of sediment within the overdeepening that could limit erosion.

(2.9) What is the influence of pre-existing (fluvial, glacial) topography on the formation, flow and erosion of large glaciers? How can corresponding transient effects be incorporated in simulations over several glacial cycles?

It has been postulated that Quaternary glacial cycles led to switches between glacial and fluvial erosion, in turn causing the landscapes to be constantly out of balance with climate. This can only be efficient if the respective glacial and fluvial erosion processes are efficient, which can take the landscape away from steady conditions. However, in places like the Alps, measured fluvial erosion rates at Holocene and present-day time scales indicate that millions of years would be required for the landform to return to a fluvial landscape. Therefore, the landscape of the Alps is clearly dominated by glacial erosion processes. Relevant to glacial erosion in the foreland is the issue of tectonics of the Alps because uplift impacts the accumulation area of the glaciers in the long term.

(2.10) Are erosion models good enough to be relied upon or would we be safer just making very conservative assumptions about how and where to bury waste? What is the status and proper role of glacier erosion modelling in addressing questions of this importance?

In relative terms, deformation and thermomechanics of glacier ice are reasonably well understood. Glacier hydrology, ice-bed interaction, basal sliding, sediment transport and interactions with permafrost are comparatively poorly understood and require considerable additional model development to be usefully incorporated into comprehensive coupled models.

Because of substantial uncertainties in the quantification of significant physical processes (e.g. sliding law, drainage physics, erosion parameterizations) as well as the relevant climate forcing of future glacial cycles (e.g. dominance of 100 ka cycle vs. 40 ka cycle, duration of current interglacial in view of anthropogenic warming, atmospheric circulation, spatial and temporal distributions and rates of precipitation) prognostic modelling is beyond our present capabilities and not recommended. In contrast, diagnostic modelling has potential value in understanding processes and quantifying uncertainties and for testing sensitivities and process parameterizations.

Modelling strategies that hold promise include ensemble modelling to explore the range of outcomes over the entire range of uncertainty in variables and extremal (end-member) modelling to bracket best and worst cases. These approaches can lead to the identification of potentially important processes and parameters and thus the component models needed (e.g. basal hydrology, sediment transport, erosion) for coupled models. To use these coupled models for prediction, they need to be validated and verified against observations for separate situations (in analogy to climate models applied to anthropogenic warming which are validated for the modern climate, verified for the last ~ 150 years using paleo-climate proxies and applied to project the next ~ 100 years). Even if models are too sensitive to many uncertain parameters to accomplish model validation and verification, they can still be used to rule out or estimate the likelihood of worst-case scenarios at specific sites.

An alternative to process-based erosion modelling is the more empirical approach of focusing on geological and observational data about erosion (e.g. looking at the ranges of observed modern glacial erosion rates in conjunction with landscape-scale denudation rates). Further, if a repository needs to be located somewhere that will not be scoured by glaciers, an obvious approach is to choose a location that will be least occupied by concentrated flows of ice and will not be affected by perturbations due to paraglacial processes (e.g. mass movements). Ice cover extent and ice flow paths can be modelled with great confidence which is probably good enough to make some meaningful predictions.

Appendix: Abstracts of presentations

This Appendix lists the abstracts (in alphabetical order) of all presentations given by the experts (see Chapter 3) on 29 April (Introduction & Context; Session 1: Measurements & Observations) and 30 April (Session 2: Modelling).

Use of low-temperature thermochronology to quantify glacial erosion

Jean Braun, Frédéric Herman, Edward J. Rhodes

Glaciers and rivers control the shape of the high relief topography of mountain ranges. However, their relative contribution in response to climatic oscillations and tectonic forcing and whether landscape can reach equilibrium conditions during the Quaternary are still unclear. Here we introduce a new thermochronometer of exceptionally low closure temperature (~ 30-35°C) based on Optically Stimulated Luminescence (OSL) dating and illustrate how it may be used to measure relief evolution and exhumation rates within the last glacial cycle in the Southern Alps of New Zealand, one of the most tectonically active orogens and an area that has experienced rapid, high magnitude climate changes. We find that exhumation rates have remained steady over the last glacial cycle and match rates observed at a million year timescale. This suggests that, despite an extreme exhumation rate of the order of 800 m in 100 ka, and the fact that in the last ~ 11-18 ka most hillslope sides have changed from U to V-shape valleys and have been dissected by debris-flows, landslides and rock avalanches, the mean exhumation rates have remained nearly constant. This may imply that tectonics, not climate, is a primary control on the rates of exhumation in tectonically active mountain belts.

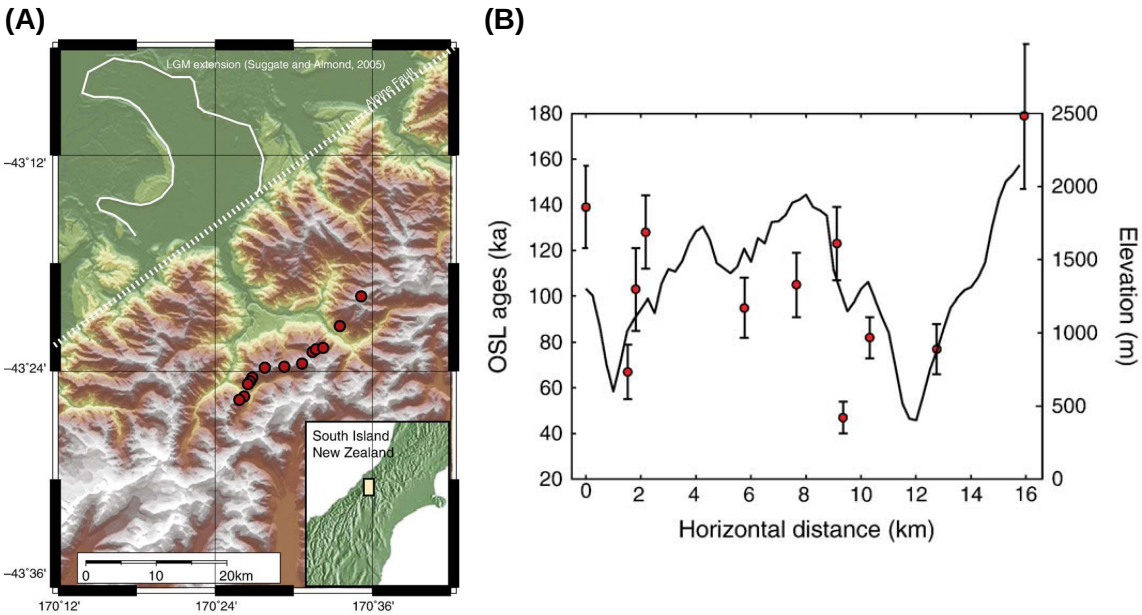

Fig. 1: Whataroa-Perth catchment area in the Southern Alps of New Zealand. (A) Red dots show the location of the analyzed samples. Inset depicts the South Island of New Zealand. (B) Ages and elevation along the profile shown in (A). Red dots are the ages and the black line depicts the topographic profile, which is interpolated along a line connecting the samples (Herman et al. 2010).

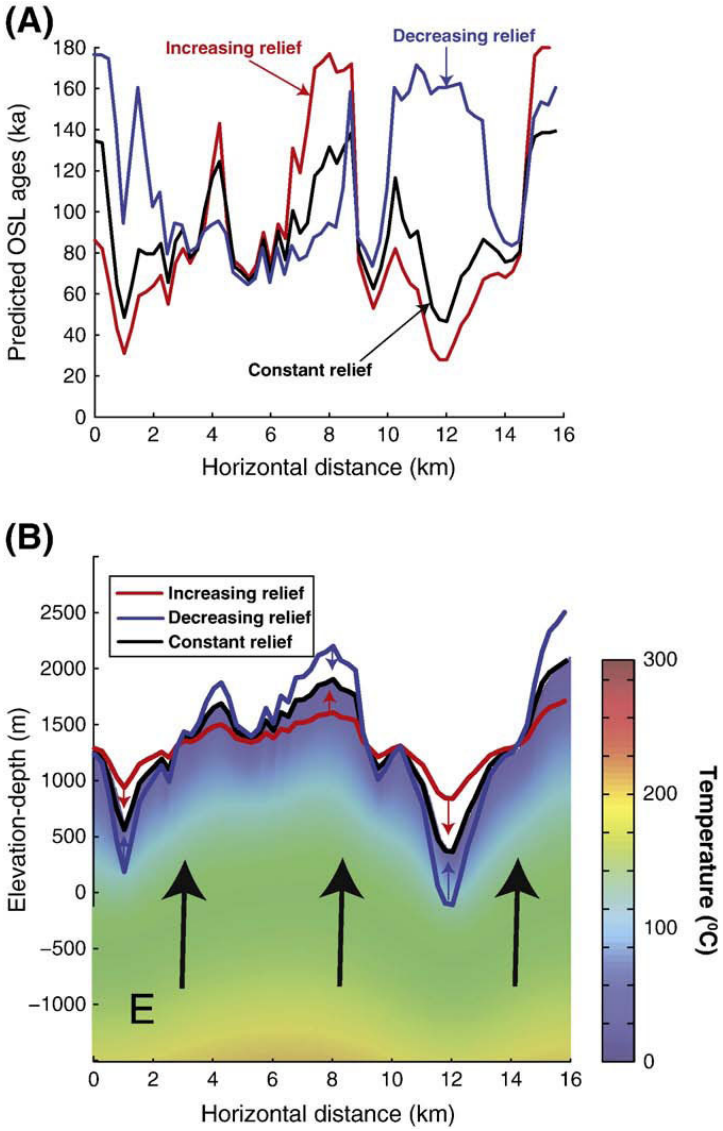

Fig. 2: OSL age predictions in response to a change of relief using the thermal model. (A) Predicted OSL ages using determined topography. (B) Prescribed change of topography and the temperature field in the crust, as predicted by the thermal model. The relief is assumed to remain steady (black line only), decrease (from blue line to black line) or increase (red line to black line) by 50 % during the last 100 ka (Herman et al. 2010).

Braun, J. (2002): Quantifying the effect of recent relief changes on age-elevation relationships. Earth and Planetary Science Letters, 200, 331-343.

Herman F., Braun, J. & Dunlap, W.J. (2007): Tectonomorphic scenarios in the Southern Alps of New Zealand. Journal of Geophysical Research, 112, B04201, doi:10.1029/2004JB003472.

Herman F., Cox, S.C. & Kamp, P.J.J. (2009): Low-temperature thermochronology and thermokinematic modeling of deformation, exhumation, and development of topography in the central Southern Alps, New Zealand. Tectonics, 28, TC5011, doi:10.1029/2008TC002367.

Herman F., Rhodes, E.J., Braun, J. & Heiniger, L. (2010): Uniform erosion rates and relief amplitude in the Southern Alps of New Zealand, as revealed from OSL-thermo-chronology. Earth and Planetary Science Letters, doi:10.1016/j.epsl.2010.06.019.

Insights into glacial landscape evolution from digital topographic analyses

Simon H. Brocklehurst

Digital topographic analyses that quantify the distinctions between fluvial and glacial landscapes allow insight into the style and scale of landscape modification attributable to glacial erosion. Quantitative comparisons between glacial and both observed and locally calibrated, simulated fluvial landscapes have been used to examine the effects of tectonics and drainage basin scale on glacial landscape modification. This topography-based approach can highlight the distribution of glacial erosion, and the factors that control glacial erosion, but since the landscape integrates processes acting over multiple glacial cycles, it cannot directly reveal the rates or timing of glacial processes.

Longitudinal profiles of drainage basins in the eastern Sierra Nevada, California, indicate the effects of ice residence time on morphology. A modest extent of glaciation results in cirques at the head of the valley, while more extensive glaciers generate characteristic steps further down the profile. However, significant modification of the profile below the long-term equilibrium line altitude (ELA) only occurs above a threshold drainage area (~ 40 km^2). In turn, a comparison of glacial longitudinal profiles from the Sierra Nevada and the more tectonically-active Nanga Parbat region of the Himalayas (Fig. 1) reveals surprisingly similar geometries for the glaciated reach, reflecting the glacial buzzsaw hypothesis (Brozovic et al. 1997), that glaciers are capable of eroding at a rate comparable with the rapid uplift, while maintaining a shallow downvalley gradient. In contrast, the erosion rate on the hillslope at the head of the valley must be much slower, since headwall relief is much greater in the Himalayan case (> 3.5 km, versus < 0.5 km), allowing the peak of Nanga Parbat to exceed 8 km.

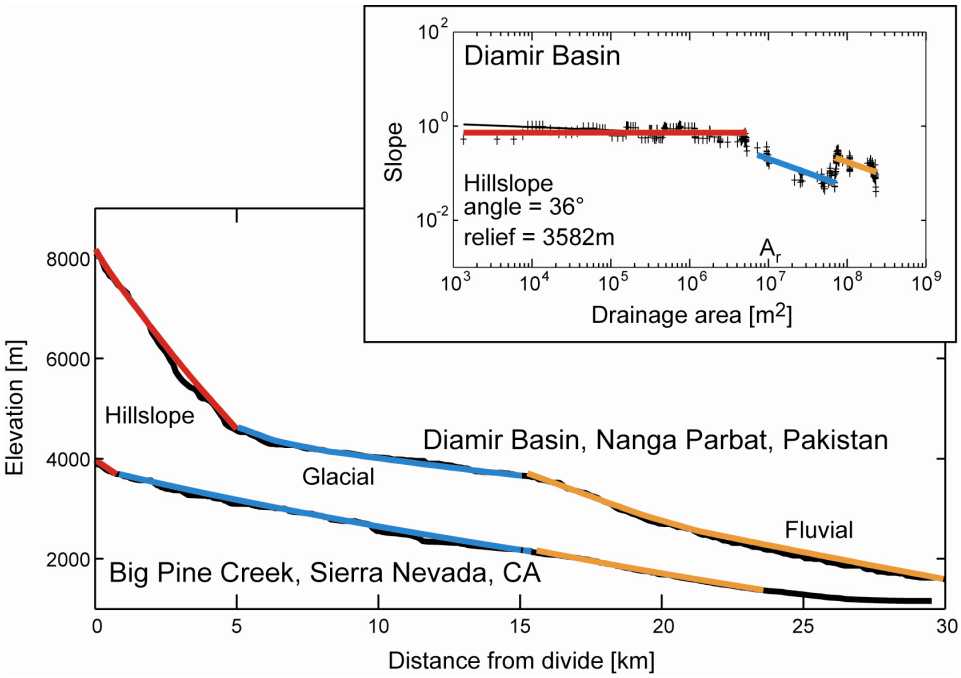

Fig. 1: Comparison of longitudinal profiles for the Diamir Basin, Nanga Parbat, Pakistan (rapid rock uplift) and Big Pine Creek, Sierra Nevada, California, USA (modest rock uplift). Glacially-dominated portion highlighted in blue (after Brocklehurst 2010).

Topographic analysis of the Teton Range, Wyoming, USA, further illustrates the importance of drainage basin scale on the evolution of a glacial landscape, and also illuminates the complex feedbacks associated with a tectonically-active range perpendicular to the prevailing winds (Fig. 2). The Teton Range is being uplifted in the footwall of the Teton Fault, which has a slip rate of ~ 2 mm/yr. The range lies at the eastern end of the Snake River Plain, and much of the precipitation reaching the range is carried by the prevailing westerly winds. Displacement on the normal fault has tilted the unconformity that caps the range, so that Precambrian basement is exposed on the eastern side, while gently westward-dipping Palaeozoic sediments drape the western side. Furthermore, the range carries a distinct asymmetry; the line through the highest peaks in the range lies to the east of the drainage divide. Drainage basins on the eastern side of the range are neatly separated into three groups: the largest basins reach beyond the high peaks to the drainage divide behind, and are characterised by shallow downvalley gradients; an intermediate set of basins drain from the high peaks with steeper gradients; and the range front features a number of shallow basins with very steep gradients. PRISM precipitation data (http://www.ocs.orst.edu/prism) and summer insolation calculations (Fig. 3) reveal the many feedbacks associated with glaciation and glacial erosion in the Teton Range (Fig. 2). High peaks focus precipitation, and represent a source of avalanching snow. Furthermore, the prevailing wind advects falling snow across the range from west to east, further than would be the case for rainfall, and also redistributes snow from the shallow, open drainage basins on the west side of the range to the more deeply-incised, sheltered cirques on the eastern side. Meanwhile, shading by the high peaks on the eastern side, plus rockfall debris from these peaks, add to the mass balance of glaciers here, promoting more extensive, longer-lived glaciers, and greater glacial incision. However, flexural modelling indicates that, despite the obvious and dramatic glacial incision in the large drainage basins, there has not been sufficient incision to drive significant isostatic uplift of the peaks (maximum ~ 150 m, only ~ 7.5 % of the range relief). This presumably reflects the narrow range width in comparison to the large sedimentary basins either side.

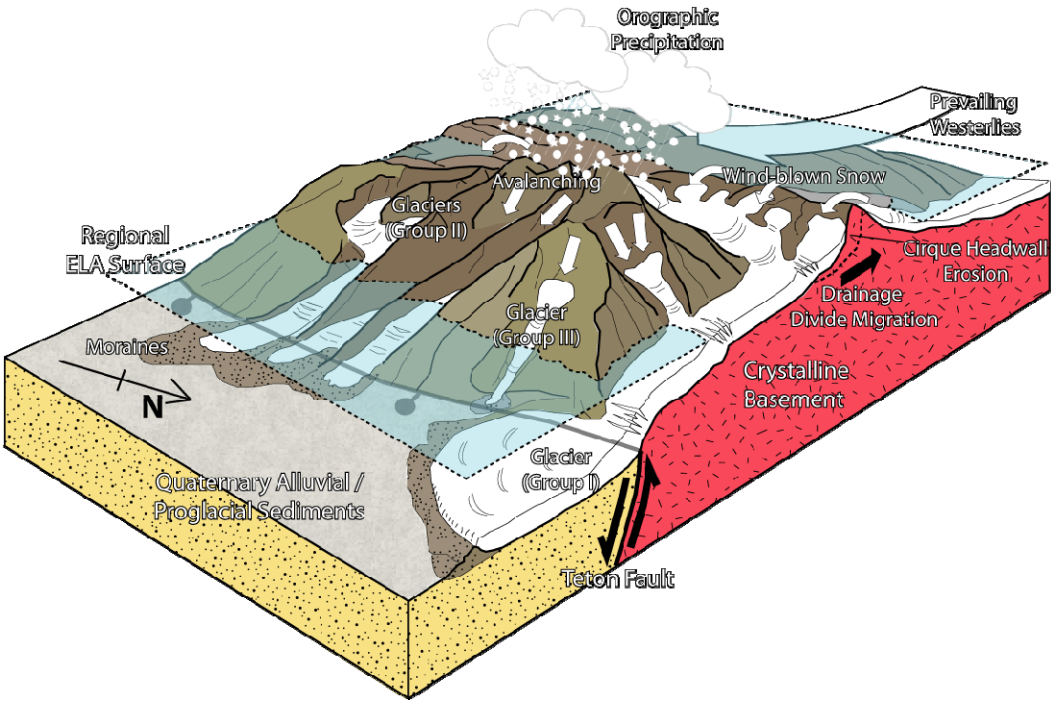

Fig. 2: Conceptual model of recent landscape evolution in the Teton Range, Wyoming, USA (after Foster et al. 2010).

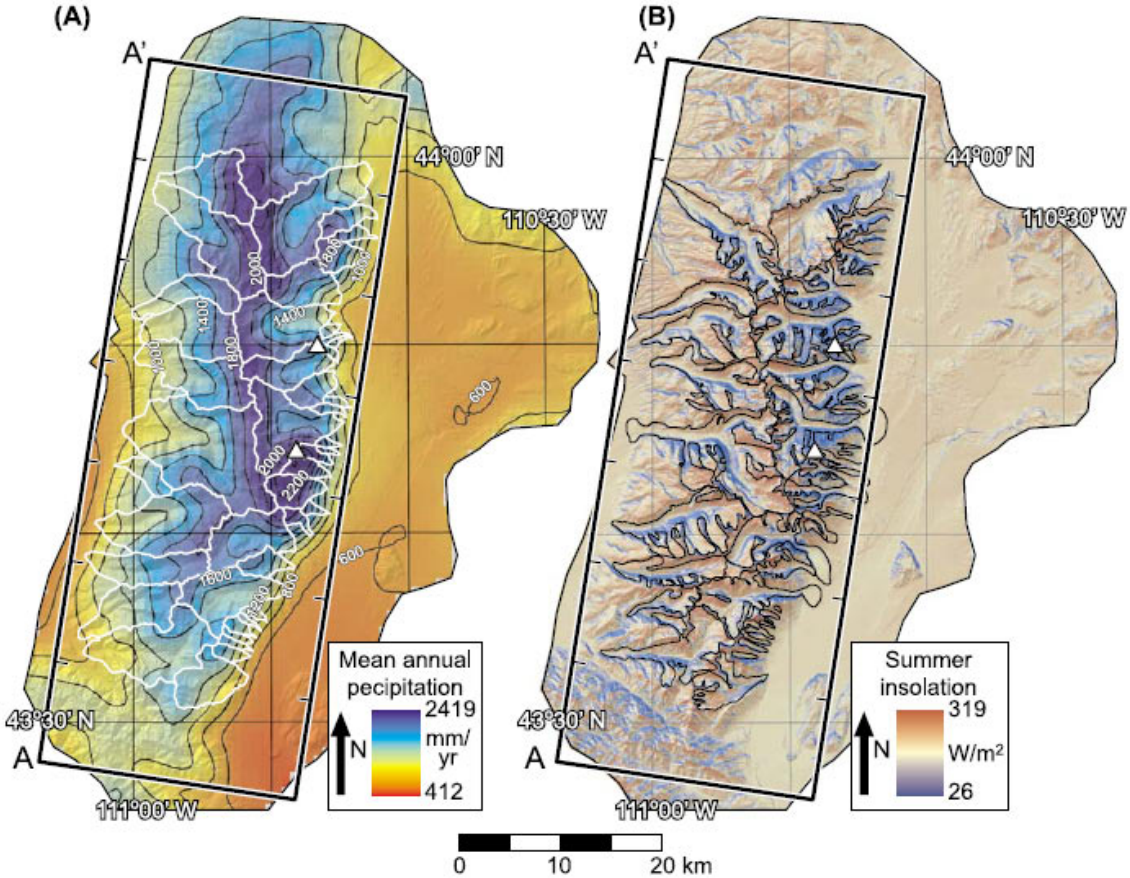

Fig. 3: Climatic parameters for the Teton Range, Wyoming, USA. (A) Mean annual (1970-2000) PRISM precipitation map. Contour interval: 200mm/yr, drainage basin outlines shown in white. (B) Summer insolation. Black outlines show reconstructed LGM glacial trimlines (after Foster et al. 2010).

The approaches outlined here have demonstrated the large range of morphologies in glacial landscapes, and also the range of influences on the development of glacial landscapes, including the importance of drainage basin size (affecting initial downvalley gradient, accumulation area, behaviour of the subglacial drainage network, ice residence and discharge, and relief), the role of the prevailing wind, and the importance of the tectonic setting.

Brocklehurst, S.H. (2010): Tectonics and Geomorphology. Progress in Physical Geography, 34(3), 357-383.

Brocklehurst, S.H. & Whipple, K.X. (2002): Glacial erosion and relief production in the Eastern Sierra Nevada, California. Geomorphology, 42, 1-24.

Brocklehurst, S.H. & Whipple, K.X. (2006): Assessing the relative efficiency of fluvial and glacial erosion through simulation of fluvial landscapes. Geomorphology, 75, 283-299.

Brozovic, N., Burbank, D.W. & Meigs, A.J. (1997): Climatic limits on landscape development in the northwestern Himalaya. Science, 276, 571-574.

Foster, D., Brocklehurst, S.H. & Gawthorpe, R.L. (2010): Glacial-topographic feeback mechanisms along normal-fault bounded ranges: the Teton Range, Wyoming, USA. Journal of Geophysical Research – Earth Surface, 115, F01007, doi:10.1029/2008JF001135.

Computer modelling of mountain glaciation

Garry K.C. Clarke, Faron S. Anslow, Alexander H. Jarosch, Valentina Radic, Christian Reuten

We have been developing numerical ice dynamics models to examine the ongoing and accelerating deglaciation of the mountainous regions of western North America (Fig. 1). The aim is to predict changes in the water cycle that will result from anthropogenically-forced climate change. The ice dynamics model assumes the shallow-ice approximation (SIA) and uses a semi-implicit method to solve vertically-integrated equations of motion at a spatial resolution of 200 m. Although the SIA can be faulted for oversimplifying the ice flow mechanics, we find that for future projections the single most important consideration is to correctly represent the mass balance forcing. We use the North American Regional Reanalysis (NARR) to represent the climate of the past 40 years and have developed and tested methods for downscaling the temperature and precipitation fields from the NARR scale of ~ 30 km to the model scale of 200 m (Fig. 2). To project glacier changes to 2100 AD we apply the delta approach to the output from standard IPPC-class GCMs driven by plausible emission scenarios (e.g., A1B). Thermal transport and debris transport can be optionally included. In addition to the conventional forward model described above, we have developed tangent linear and adjoint models which enable model sensitivities to be systematically examined.

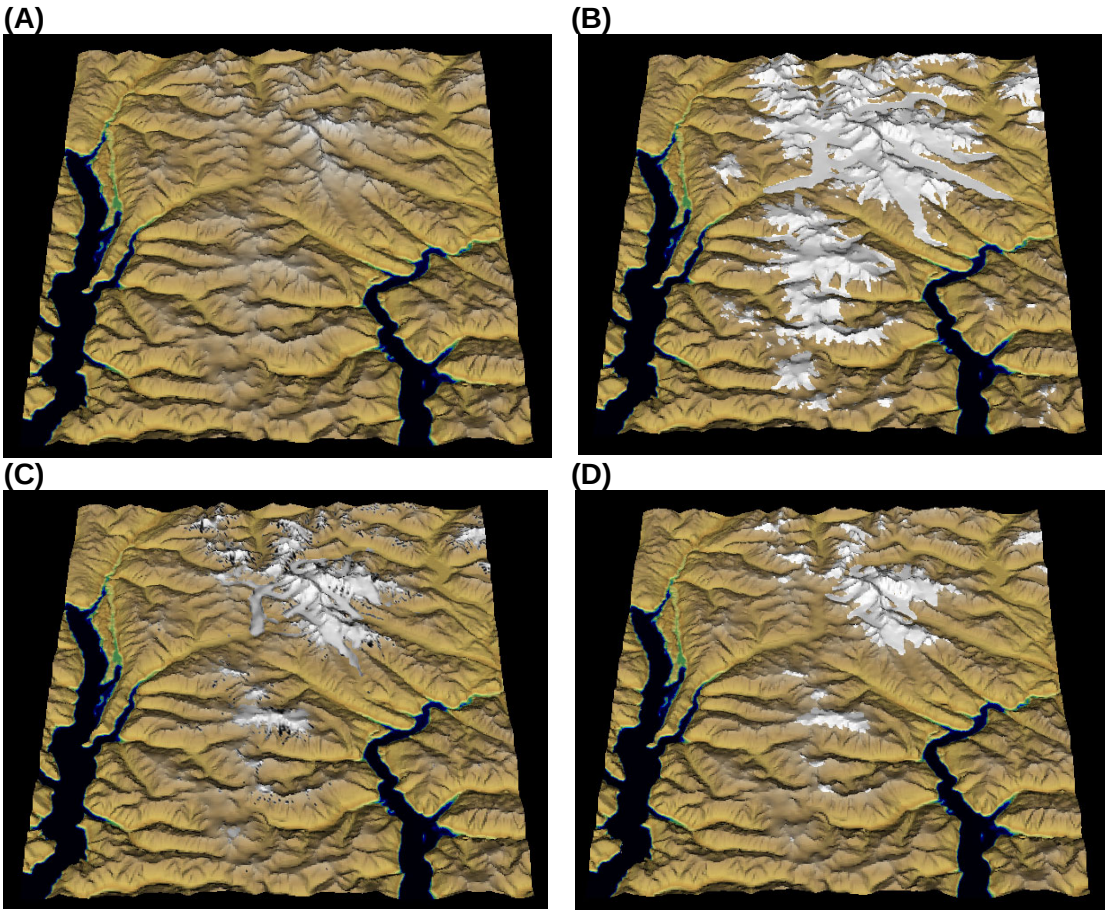

Fig. 1: Results from model calculations using Mt. Waddington, BC as a test example. (A) Subglacial topography. Extent of glaciation (B) today and in 500 years (C) with and (D) without inclusion of debris transport.

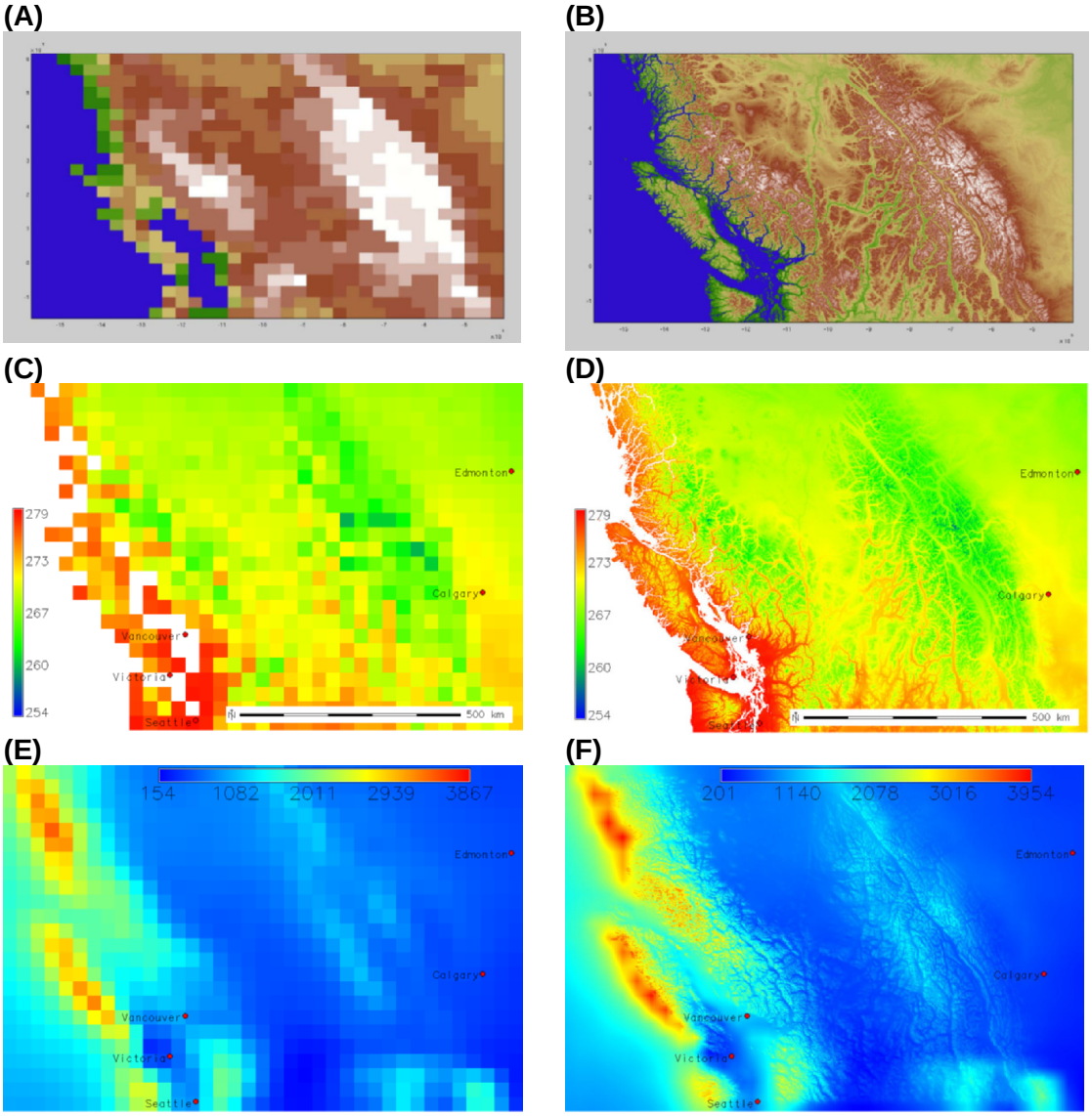

Fig. 2: Downscaling of the (A), (B) topography, (C), (D) temperature field and (E), (F) precipitation field in southwestern Canada from the North American Regional Reanalysis (NARR) scale of ~ 30 km (left panels) to the model scale of 200 m (right panels).

Effects of water pressure fluctuations on quarrying: subglacial experiments using acoustic emissions

Denis O.H. Cohen, Neal R. Iverson

Quarrying, the growth and coalescence of cracks in subglacial bedrock and dislodgement of resultant rock fragments, is probably the most important mechanism of glacial erosion but may be one of the least understood subglacial processes. Although evidence indicates that erosion rates depend on sliding speed, rates of crack growth in bedrock may be enhanced by changing stresses on the bed caused by fluctuating basal water pressure in zones of ice-bed separation. To study quarrying in real time, a 12 cm high granite step with a crack in its stoss surface was installed under 210 m of ice at the bed of Engabreen, a temperate glacier in northern Norway (Fig. 1). Acoustic-emission sensors monitored crack-growth events in the step as ice slid over it. Vertical stresses on the step and bed, water pressure, and cavity height in the lee of the step were also measured. Water was pumped under high pressure to the lee of the step several times over eight days. Pumping initially caused opening of a leeward cavity, which then closed after pumping was stopped and water pressure decreased. During closure of the cavity, acoustic emissions emanating from the base of the crack in the step increased dramatically (Figs. 2 and 3). With repeated pump tests, this crack grew with time until the step's lee surface was quarried (Fig. 4). Fluctuating water pressure may commonly be necessary to exceed stress thresholds required for crack growth. Stress changes on the bed due to water-pressure fluctuations will increase in magnitude and duration with cavity size, which may help explain the effect of sliding speed on erosion rates. Ultimately, rates of quarrying, and hence erosion, may depend on the magnitude and frequency of water pressure fluctuations.

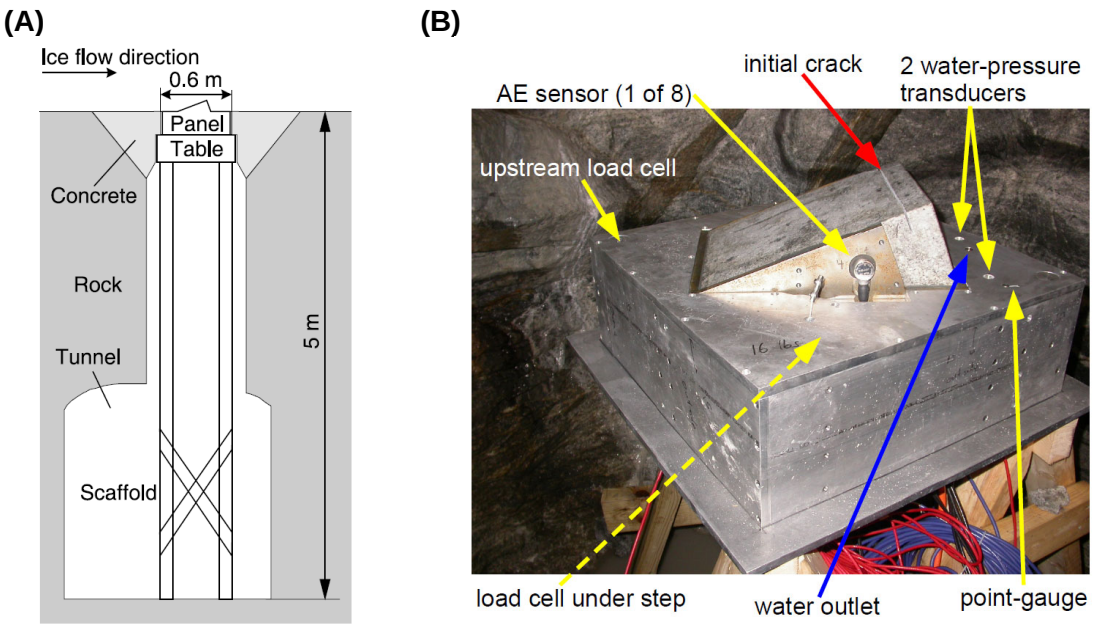

Fig. 1: Experimental setup beneath 210 m of ice at the bed of Engabreen, Norway. (A) Cross section of tunnel and vertical shaft showing panel, supporting table and scaffold. (B) Panel on table at the bottom of vertical shaft (Cohen et al. 2006).

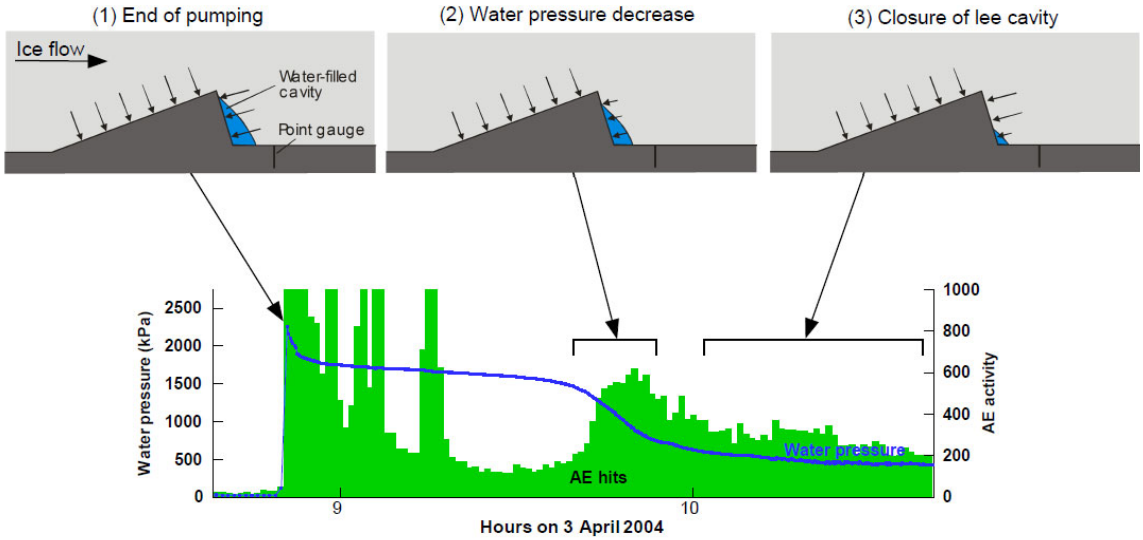

Fig. 2: Average water pressure and acoustic emission (AE) frequency during a pump test. AE activity is high during pumping because of panel vibrations.

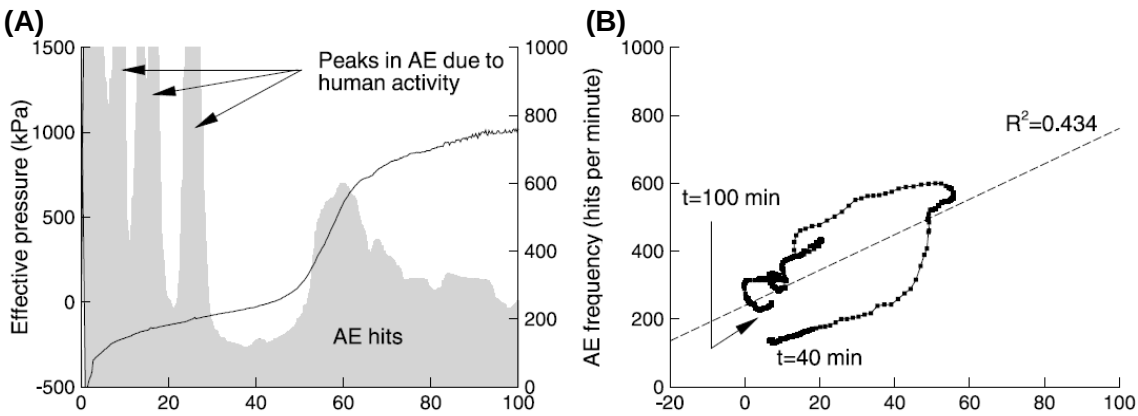

Fig. 3: (A) Effective pressure and acoustic emission (AE) frequency as a function of time during a pump test. (B) Corresponding plot of AE frequency as a function of the time derivative of the effective pressure. Dashed line indicates linear best fit. *R* is the coefficient of linear regression and *t* is time since the start of the pump test (Cohen et al. 2006).

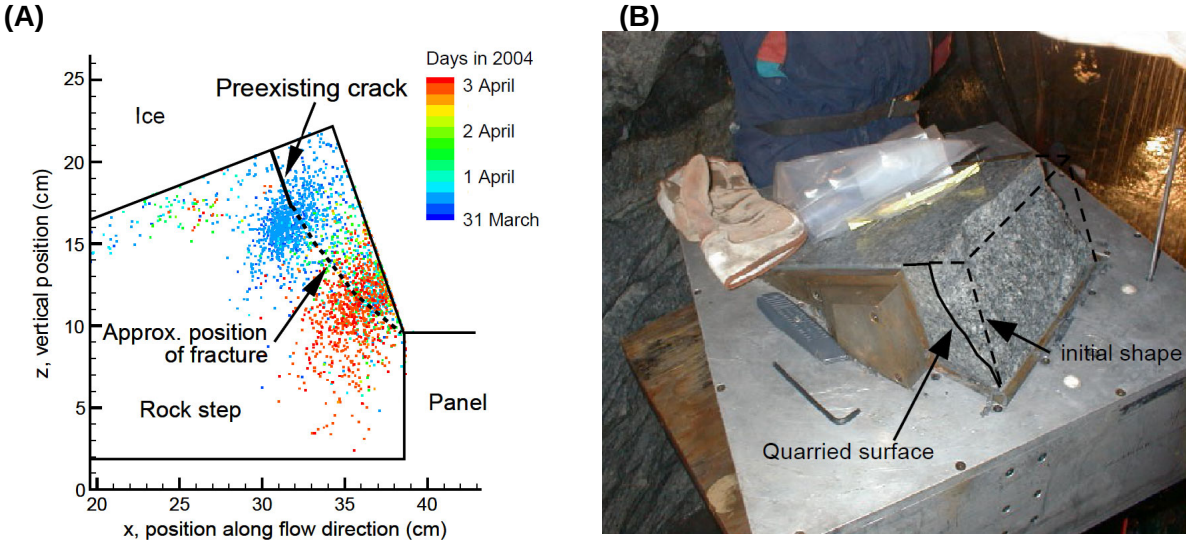

Fig. 4: (A) Location of acoustic emission (AE) events in the plane parallel to ice flow during the pump tests. Each square indicates an event captured by all eight sensors. Colour indicates timing of event. Outlines of granite step, panel, preexisting crack and fracture made during the experiment are also shown. (B) Step at end of experiment with quarried lee surface (Cohen et al. 2006).

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Modelling glacial erosion with a higher order ice-sheet model

David L. Egholm

The rugged topography of mountain ranges represents a special challenge to computational ice-sheet models simulating past or present glaciations. Topographic relief steers glaciers through relatively narrow and steep valleys, and as a consequence hereof, the flow rate of alpine-style glaciers varies significantly at length scales comparable to that of the topography. Localized flow rate undulations generate longitudinal and transverse stress gradients within the ice, which, in turn, are of known importance to the flow itself, whether by internal deformation of the ice or by basal sliding, and to the interaction with topography through glacial erosion.

In the presentation, I outline the pros and cons of a new model framework which, on the one hand, takes into account the 'higher-order' effects related to steep and rugged bed topography while it still, on the other hand, provides the computational efficiency needed for three-dimensional simulations of glaciation and landscape evolution in response to e.g. long-term climate variations. As such, the new model framework (iSOSIA) is well suited for computational experiments exploring long-term feedbacks between climate, orography, and plate tectonics.

Shallow ice approximations are computationally efficient, primarily because ice flow can be treated in a depth-integrated manner. This effectively reduces the dimensionality of a flow problem from three to two, and the computational cost of solving the flow equations is lowered dramatically. Hence, with a shallow ice approximation it becomes possible to simulate iteratively the temporal evolution of glaciers under e.g. climatically varying conditions or under the influence of topographic feedbacks associated with glacial erosion. However, the accuracy of shallow ice approximations depends on a small aspect ratio between the scale of ice thickness and length scales of horizontal variations in ice thickness, bed topography or ice flow velocity (Hutter 1983).

As explained by Hutter (1983), the shallow ice approximation is an asymptotic theory, which can be formulated in different orders with varying demands to the magnitude of the aspect ratio mentioned above. The simplest and most popular shallow ice approximation includes only zero'th order terms and therefore requires the aspect ratio to be very small (for example, bed slopes should not exceed 0.01). This condition is violated for glaciers in areas of increased topographic relief, or even for an ice-sheet close to its margin, where the bed may be flat, but the ice surface is characterized by steep gradients due to large ice thickness variations. This conventional zero'th order shallow ice approximation is traditionally referred to in the literature as 'the Shallow Ice Approximation' (SIA).

Many higher-order approximations exist (e.g. Hindmarsh 2004; Pattyn et al. 2008) that provide increased accuracy in situations where the aspect ratio is larger. In most cases the higher-order approximations include the longitudinal stress components (*membrane stresses* (Hindmarsh 2006)) and their spatial gradients, which seems sufficient in many 'ice-sheet scenarios' where the ice surface is only gently sloping (Pattyn et al. 2008). However, as described by Baral et al. (2001), the *membrane stresses* only partially represent the higher-order terms of the general shallow ice approximation.

With iSOSIA, we attempt to follow a mathematically consistent path in implementing a depth-integrated version (iSOSIA) of the full Second-Order Shallow Ice Approximation (SOSIA) (Baral et al. 2001). This allows the aspect ratio to be significantly larger (at least up to 1) by including all zero'th, first, and second order terms of the shallow ice approximation.

Besides increasing the confidence in the numerical solution, the higher-order terms of iSOSIA allows us to simulate ice flow in steep terrains at catchment scale, using relatively high spatial and temporal resolution. To study the denudation of mountain ranges, it is essential that we can model the interactions between glaciers, sediments, and landscapes on a catchment scale where solid observations relating to landscape morphology and denudation patterns exist.

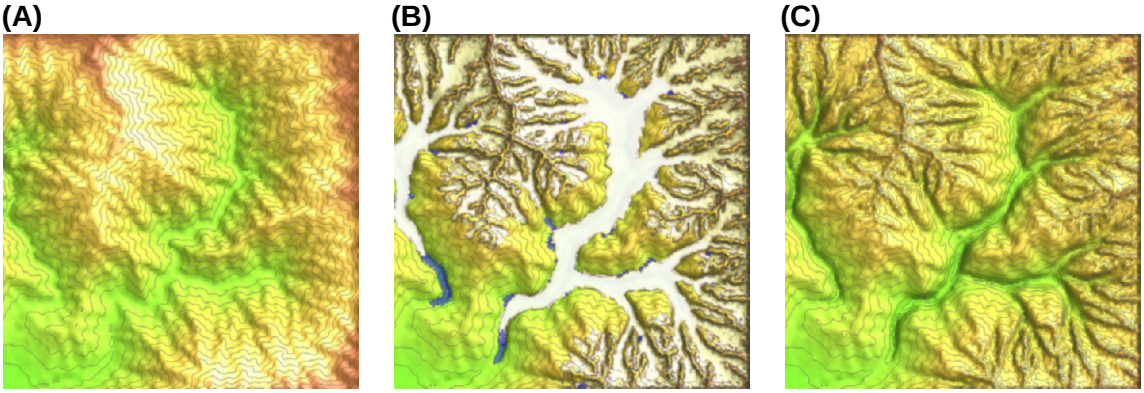

Fig. 1: An example of modelled glacial erosion in an initially fluvial model landscape. (A) The initial fluvial landscape. (B) The glacially eroded landscape with ice present. (C) The eroded landscape without ice. Glacial erosion occurs in response to basal sliding, which in turn is a function of the basal shear stress. The higher-order terms of iSOSIA have a marked impact on the distribution and magnitude of basal shear stress. Glacial landforms, such as valley overdeepenings, U-shaped and hanging valleys, cirque valleys and arêtes, develop naturally as a consequence of feedbacks between topography and basal shear stress.

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The influence of various periglacial and glacial processes in an alpine glacial erosion model

Todd A. Ehlers, Brian J. Yanites

The evolution of mountain topography and sediment fluxes to adjacent basins is dictated by variations in rates of rock-uplift, climate, lithology, and vegetation. In this study, we investigate transients in mountain erosion and morphology over million year time scales due to glacial-interglacial cycles imposed on landscapes previously dominated by fluvial and hillslope processes.

In our approach, we couple a modified shallow ice approximation glacial model (Herman & Braun 2008) to a fluvial and hillslope surface process model (Braun & Sambridge 1997). The models were modified to account for orographic precipitation, positive-degree-day melting mass balance, rock and snow avalanching, frost cracking and diffusion based periglacial processes, ice buoyancy and calving, super freezing at the glacier base, and instantaneous glacial sediment transport. Two types of model simulations are presented: (i) A calibration of the glacial erosion model to mineral cooling ages preserved in moraines from the Sierra Nevada Mountains, California. And (ii) a sensitivity analysis of glacial, fluvial, and hillslope erosion rates and magnitudes starting with an equilibrium (synthetic) fluvial landscape that is subjected to repeated glacial cycles. We use an equilibrium fluvial landscape generated with rock uplift rates between 0.25-1.0 mm/yr. The landscapes are then subjected to repeated glacial cycles of different periodicity and intensity. Variations in predicted glacial basal sliding velocity, erosion, topography and sediment flux are tracked (e.g. Fig. 1).

Results are as follows: (i) Model predicted ice thicknesses and sliding velocity are highly sensitive to whether uniform or orographic precipitation is considered in the simulation. Not considering an orographic distribution of snowfall results in an under prediction of erosion for the same mean precipitation rate as the uniform precipitation approach. (ii) Snow avalanching from ridges to the glacier has a significant impact on the glacier mass balance and predicted patterns of glacial erosion. Including snow avalanching in simulations results in thicker and more erosive glaciers as well as exposing more hillslope area to periglacial processes. (iii) A modified shallow ice approximation model accurately represents the pattern of glacial erosion recorded in detrital thermochronometer data from the Sierra Nevada Mountains, but only for a range of constriction factors ($K_c \cong 10^1$-10^3). (iv) Results indicate that glacial processes increase rates of valley bottom erosion by a factor of two or higher than fluvial processes, a result consistent with low-temperature thermochronological data from a number of glaciated catchments. Increased rates of hillslope and ridgetop erosion during both glacial and interglacial times occur in response to increased glacial valley erosion. And (v) for simulations considering glacial erosion over multiple glacial-interglacial cycles a lag time of ~ 1 Myr is observed between the onset of the first glacial cycle and the predicted maximum in glacial erosion rates.

Taken together these results demonstrate the utility in using detrital thermochronometer data to calibrate landscape evolution models. Furthermore, we find that glacial erosion over multiple glacial-interglacial cycles is highly variable with maximum erosion being reached only after several cycles.

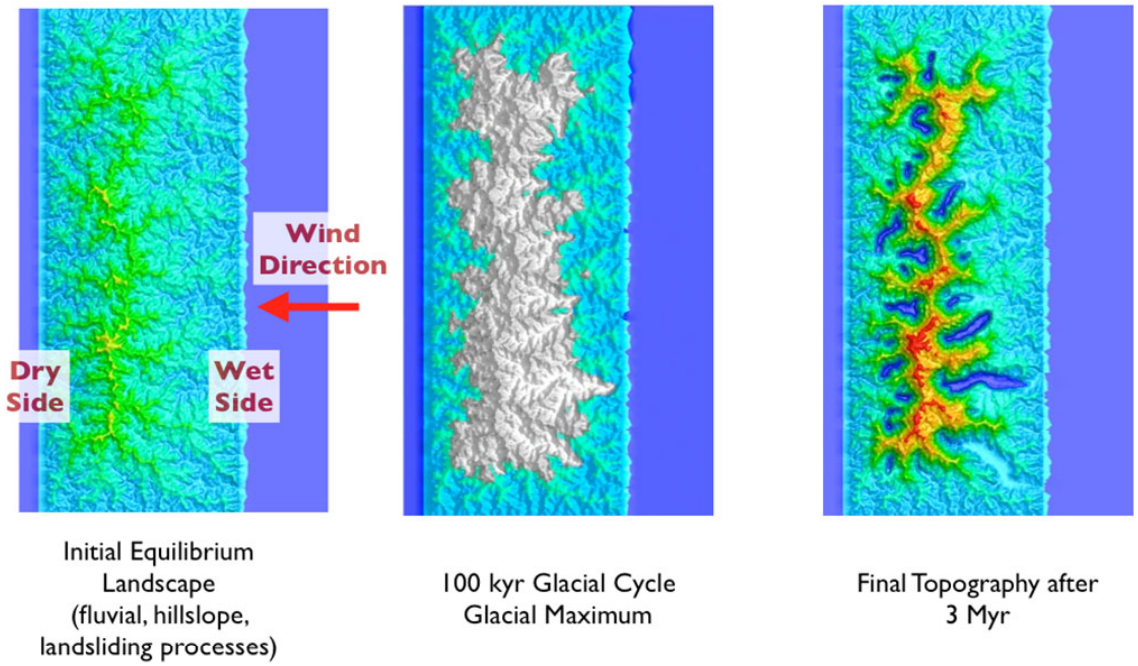

Fig. 1: Example of a coupled orographic precipitation and glacial, fluvial, and hillslope erosion model. Simulations were conducted on an initial equilibrium landscape generated from fluvial and hillslope processes with a rock uplift rate of 0.5 mm/yr. The landscape is then subjected to repeated glacial-interglacial cycles and variations in erosion by different geomorphic processes are tracked. Models are capable after multiple glacial cycles of producing glacial over deepenings (right panel). The width and length of mountain topography in the above is 150 and 400 km, respectively.

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Quantifying glacial bedrock erosion with cosmogenic nuclides

Derek Fabel

The erosive capacity of glaciers is an inductively reasoned conclusion based on observational evidence of characteristic glacial landforms such as over-deepened valleys. While theoretical models have provided valuable information on the potential mechanisms of glacial erosion, they cannot, as yet, prove their predictions because boundary conditions are not well enough known. Testing the models by direct field observations is difficult, and rates of glacial erosion have as yet not been directly measured. Nonetheless, the view that glaciers are very erosive is generally accepted within the earth sciences.

Quantification of average glacial erosion rates has been made possible with the advent of reliable measurements of in situ cosmogenic nuclide concentrations in glacially abraded bedrock surfaces. For glaciers to remove inherited cosmogenic nuclides, accumulated in bedrock as a result of exposure to cosmic rays during an earlier ice-free period, requires more than 2.5 metres of bedrock erosion. Using cosmogenic nuclides it is possible to determine if there has been less than 2.5 m of bedrock erosion, but it is not possible to place limits on how much more material has been eroded.

Measurement of cosmogenic nuclide concentrations in bedrock surfaces of formerly glaciated areas has shown that glacial erosion is spatially restricted at ice sheet and valley scales. In East Antarctica, striated bedrock surfaces yield nuclide concentrations in excess of what would be expected if the overriding ice had eroded sufficient bedrock. Similarly, apparent cosmogenic surface exposure ages of bedrock at the centre of the Scandinavian ice sheet are often well in excess of the deglaciation age expected if the ice had been erosive. This is not only manifested on interfluves, away from areas of former ice discharge, but also locally in major trunk valleys. Low bedrock weathering and erosion rates in regions formerly occupied by ice sheets are probably due to frozen bed conditions at the base of ice sheets.

Results from transects across U-shaped valleys in mid-latitude mountains indicate that glacial bedrock erosion in excess of 2.5 metres is confined to the valley centre and decreases significantly up the valley sides (Fig. 1). Hence glacial bedrock erosion deepens but does not necessarily widen valleys. Low bedrock erosion rates along U-shaped valley sides of temperate alpine glaciers may reflect optimisation of the valley form during earlier glaciations.

The broad implication of these cosmogenic nuclide measurements is that contrary to popular opinion, glaciological processes, at least during the last glacial cycle, have not generally been efficient agents of bedrock erosion. Yet the physical landscape evidence suggests considerable bedrock erosion during glacial cycles, suggesting that multiple glaciations are required to form over-deepened valleys. Dating the formation of over-deepenings remains an unresolved problem.

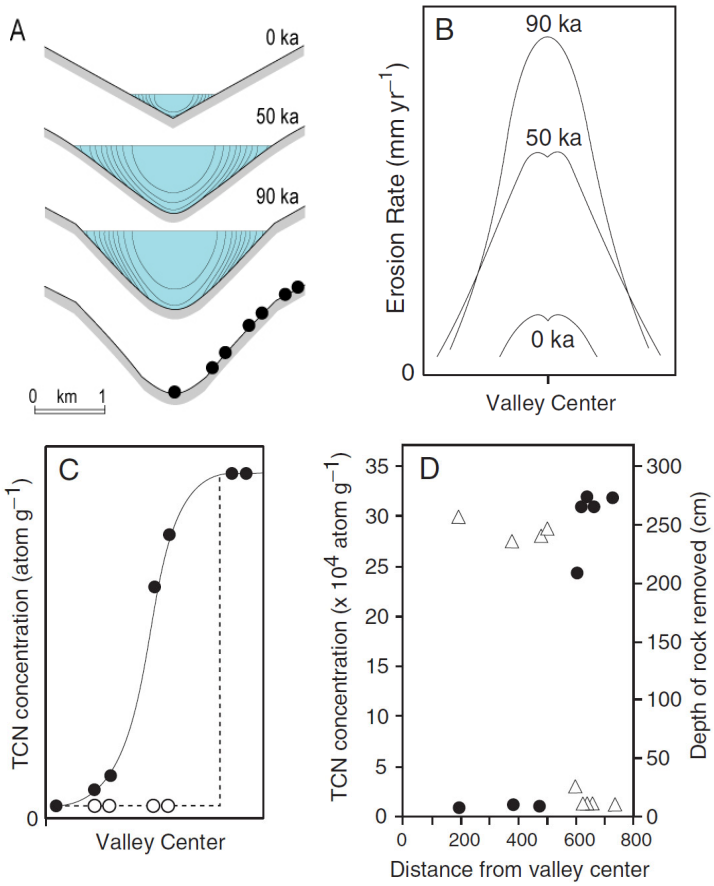

Fig. 1: (A) Simulation of cross-section form development over an idealized 100 ka glacial cycle with temporally variable ice discharge. Velocity contours for the glacier sections are in units of 10 percent of the maximum velocity for the section with the central contour in each case at 90 percent. Dots mark hypothetical sample locations shown in (C). (B) Predicted erosion rate plotted for different times during the simulation indicate that erosion decreases toward the lateral limit of the glacier. (C) Hypothetical terrestrial cosmogenic nuclide (TCN) concentrations in samples collected along a valley transect (A) for the case where deep glacial erosion has completely removed the TCN inventory (open circles) and the case of TCN inheritance due to insufficient glacial erosion (dots). (D) TCN inheritance values (dots) and calculated depths of glacial erosion (triangles) for data from Sinks Canyon,Wyoming (Fabel et al. 2004).

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Progress in modelling glacier and ice-sheet hydrology

Gwenn E. Flowers

Hydrology is an important control on erosion and sediment transport processes in subglacial and periglacial environments. Realistic models of glacier hydrology are therefore in demand, particularly to improve our characterization of subglacial water pressures and flow velocities. These quantities have relevance to ice dynamics, and to erosion and transport processes directly. Our early attempts to model glacier hydrology focused on a very simple but comprehensive two-dimensional treatment of the system that parameterizes morphological transitions in the basal drainage system (e.g. Flowers & Clarke 2002a, b), but does not properly account for the physics of efficient (channelized) subglacial drainage. Our recent efforts (Pimentel & Flowers 2010; Pimentel et al. 2010) have been focused on the development of a higher-order flowband model of ice dynamics, a parameterization of lateral drag that permits sliding on the valley side walls, and the incorporation of a Coulomb-friction law to describe the basal boundary condition. This model is coupled to a two-component model of subglacial hydrology describing an interacting sheet ("slow" or "distributed" system) and ice-walled conduits ("fast" or "channelized" system) (e.g. Flowers 2008, Fig. 1). The basal water pressure determined by this model is used to calculate effective pressure for the Coulomb friction law. Coupled simulations of a hypothetical seasonal transition qualitatively exhibit many features observed in the field (Fig. 2): as surface meltwater impinges on the slow basal drainage system, basal water pressures increase, reducing the basal shear stress and producing higher basal flow rates; the reduction in basal traction is accommodated through an increase in lateral drag and longitudinal stress gradients. With sufficient water input, conduits develop out of the sheet and ultimately accommodate most of the discharge, relieving the high basal water pressures and reducing basal flow rates (Pimentel & Flowers 2010). This model has been used to explore aspects of ice-sheet drainage for Greenland outlet glaciers, but would require modification to be applied to a broader range of ice-sheet problems. Further developments should address the thermodynamics of the full system (basal hydrology and coupling to ice dynamics) and the extension to two- and three-dimensions.

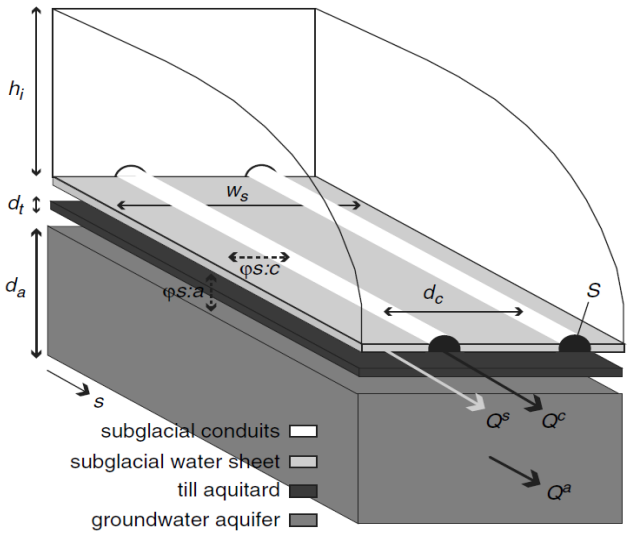

Fig. 1: Conceptual model of drainage system geometry, including 'fast' subglacial drainage (conduits), 'slow' subglacial drainage (water sheet), and groundwater flow (confined aquifer). Annotations: ice thickness h_i, aquitard thickness d_t, aquifer thickness d_a, flowline coordinate s, characteristic subglacial sheet width w_s, conduit spacing d_c, conduit cross-sectional area S, water-sheet discharge Q^s, conduit discharge Q^c, aquifer discharge Q^a (Flowers 2008).

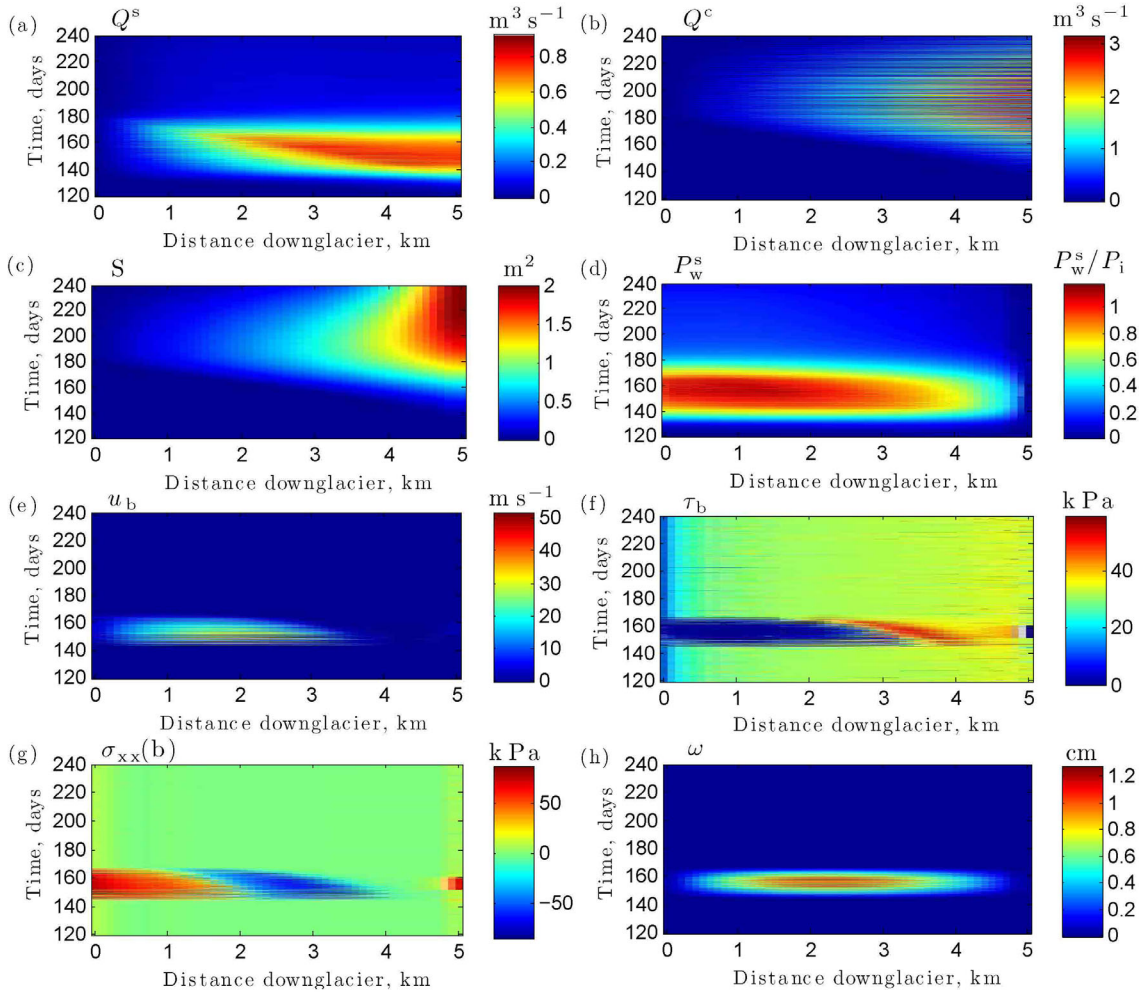

Fig. 2: Position–time diagrams for key model variables (values coloured according to the scale bars). (a) Simulated water-sheet discharge Q^s. (b) Conduit system discharge Q^c. (c) Conduit cross-sectional Area S. (d) Basal water pressure P_w^s (normalized to P_i, where P_i is the ice-overburden pressure). (e) Basal flow speed u_b. (f) Basal shear stress τ_b. (g) Longitudinal stress σ_{xx}. (h) Vertical uplift ω.

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A new method to estimate glacial erosion rates using *in situ* ^{14}C and ^{10}Be: an example from the Rhone Glacier

Brent M. Goehring, Joerg Schaefer, Christian Schlüchter, Naki Akçar, Nathaniel Lifton, Robert Finkel, Richard Alley

Fundamental to understanding glacier ice-bed interface dynamics is the spatial distribution of glacial erosion rates via abrasion and plucking. We present initial efforts to directly measure glacial abrasion rates using *in situ* cosmogenically produced ^{14}C and ^{10}Be from proglacial bedrock exposed by the retreating Rhone Glacier, Switzerland (Fig. 1). Our results indicate a systematic increase in glacial abrasion rates towards the center of the glacial trough (Fig. 2), as is expected to occur with the observed increase in ice velocity away from the lateral ice margins. Abrasion rates range from 0.12 to 1.9 mm yr^{-1}. These values are generally less than rates determined from sediment load estimates from other alpine glaciers; however, this could be due to bias of our method towards abrasion, rather than abrasion and plucking. Results from this study and future studies will allow for evaluation of existing glacier erosion theory, as well as refine models of sediment production in high-order glacier and ice sheet models.

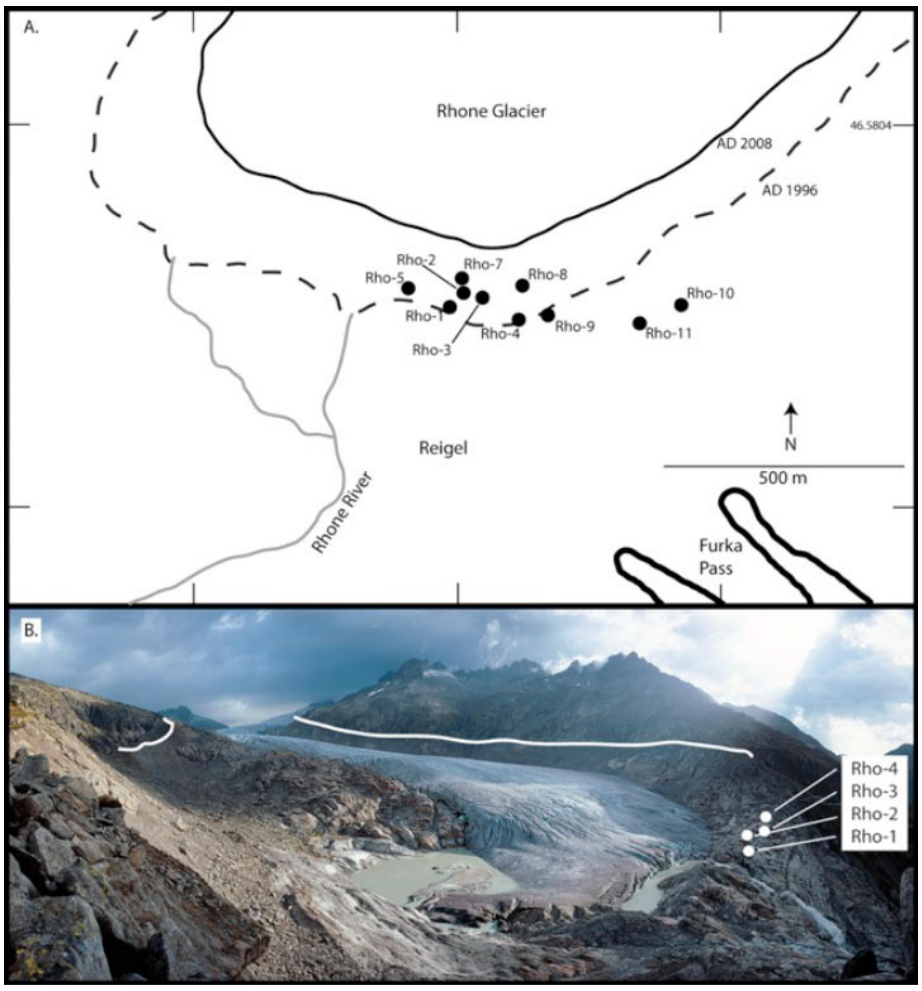

Fig. 1: Map and photograph of Rhone Glacier, Switzerland showing the proglacial bedrock sample locations.

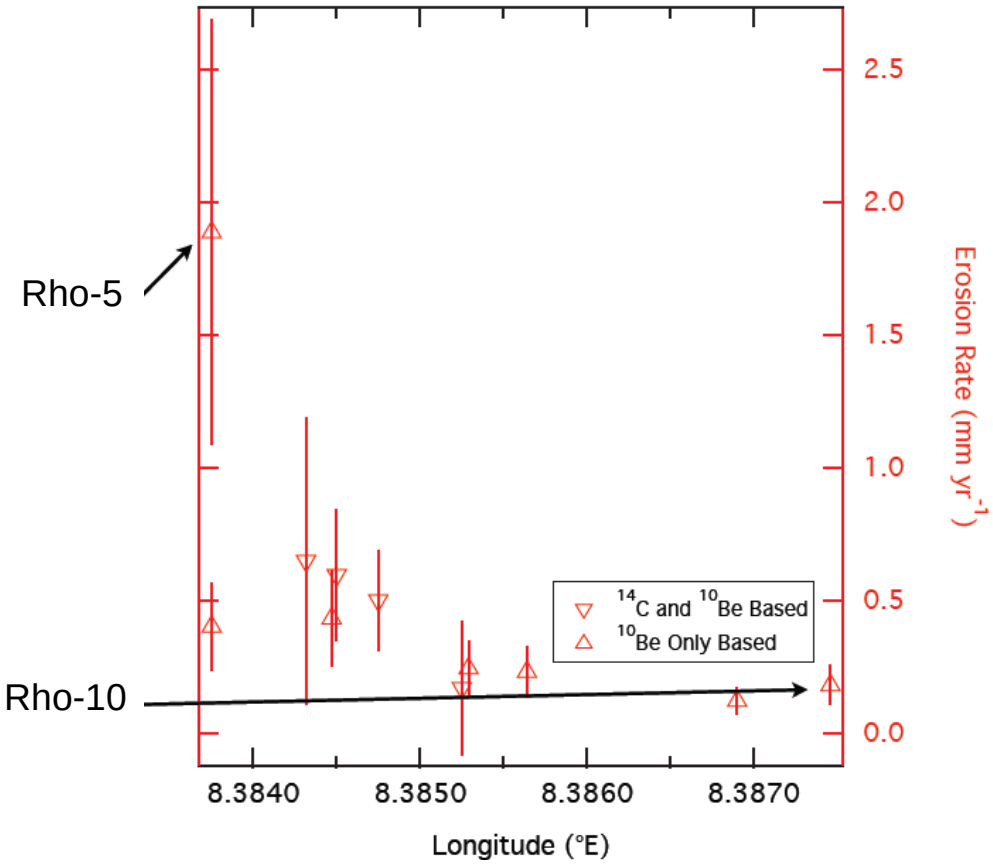

Fig. 2: Glacial erosion rates as a function of distance from the center of the glacial trough.

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Rock and sediment beds of mountain glaciers – an alternative (quantitative) approach

Wilfried Haeberli

As glaciers constitute a transfer of rock particles, the ratio between input and output of debris primarily decides on sediment fluxes within glacier systems and on the characteristics of glacier beds. A corresponding index for predicting rock and sediment beds of mountain glaciers (Fig. 1) was introduced by Haeberli (1986, 1996), later applied to glacier inventory data (Maisch et al. 1999) and implemented in a GIS (Fig. 2, Zemp et al. 2005). In its simplest ("light") version (optimally suitable for field observation and interpretation) it relates rock-wall height (for input) with the transport capacity of the meltwater stream (for output) given by the product of mean precipitation and glacier area (runoff) times the average inclination of the meltwater stream leaving the glacier margin. The approach is calibrated by observed conditions in glacier forefields exposed by ice retreat since the Little Ice Age and appears to be quite robust and realistic. It predicts mixed rock/sediment beds for LGM ice conditions in northern Switzerland as both, debris input from the distant, relatively small rock faces to the enormous and high-reaching ice masses, as well as sediment evacuation on flat terrain and by dry-climate meltwater streams must have been limited. Such considerations neglect the uptake of sediments at the glacier bed – a process, which could increase in importance with growing glacier size; they nevertheless indicate that the entire glacial/periglacial system rather than an isolated subglacial process should be taken into account and that the evacuation of sediments beyond the glacier margin, i.e. the activity of sub- and proglacial meltwater, plays a key role. With the slope of the proglacial terrain becoming zero or even inverse, lake formation builds up efficient sediment traps – a rather common feature of Ice Age glaciers in northern Switzerland. The index is now applied to modern glacier inventory data and digital terrain information in combination with modelling overdeepened parts of present-day glacier beds and possible sites of future lake formation in deglaciating high-mountain regions in view of managing such future lakes with respect to natural hazards (floods, debris flows) and hydropower development potential (cf. Frey et al. 2009, Linsbauer et al. 2009).

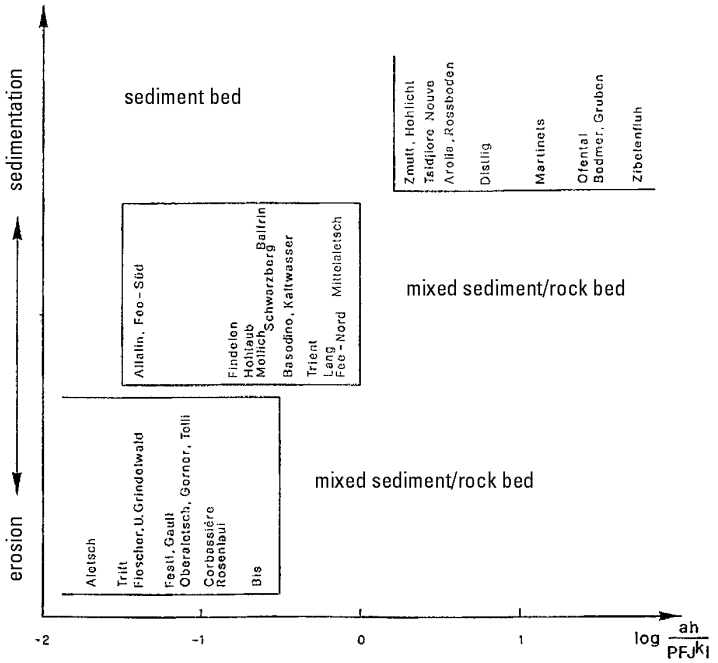

Fig. 1: Characteristics of the glacier bed as a function of an erosion-/sedimentation index expressed as the ratio of debris input (rock-wall height *h*, weighting factor for basal vs. supraglacial debris transport *a*) to debris evacuation (precipitation *P*, glacier surface area *F*, slope of proglacial stream *J*, hydrodynamic constant *k*, glacier length *l*) (Haeberli 1986).

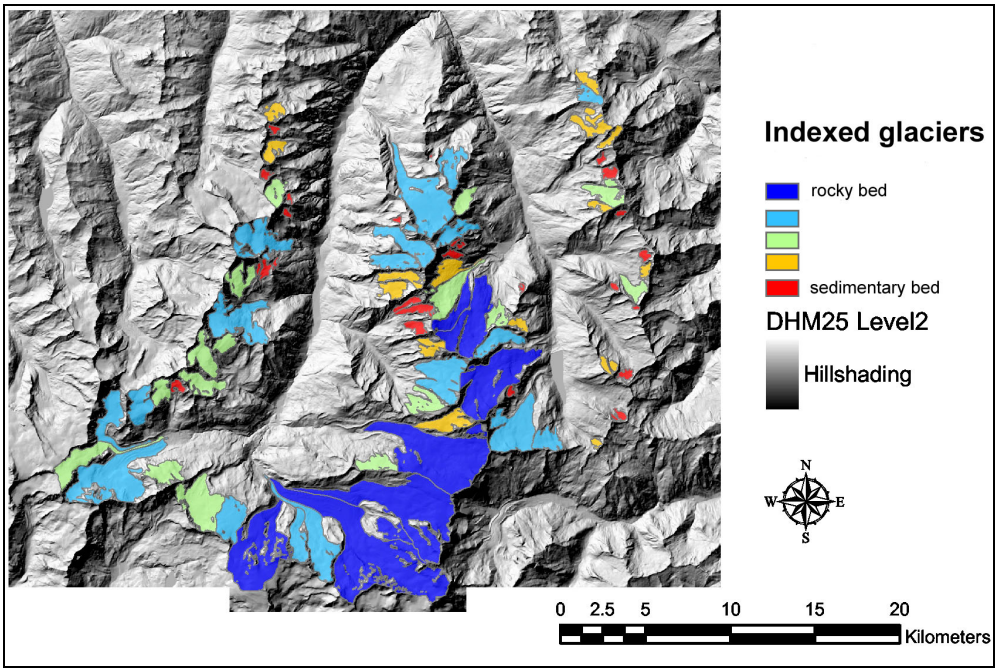

Fig. 2: Indexed glaciers in the Valais, Switzerland. Indexed glaciers are shown in five classes going from rocky beds (blue) towards sedimentary beds (red) (Zemp et al. 2005).

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Glaciological conditions in northern Switzerland during recent ice ages

Wilfried Haeberli

Based primarily on reconstructions and various model calculations, quantitative estimates are compiled concerning ice conditions in northern Switzerland during the past Ice Age. The penetration of winter sea ice to low latitudes and the corresponding closure of the Atlantic Ocean as a humidity source caused extremely cold/dry conditions in central Europe during the time period of maximum cold and most extended area of surface ice. At this stage, the large lobes of the piedmont glaciers spreading out over much of the Swiss Plateau were predominantly polythermal to cold, surrounded by continuous periglacial permafrost up to 150 m thick and characterised by low driving stresses (typically 30 to 50 kPa). Mass balance gradients, mass turnover and ice flow velocities on these piedmont glaciers were correspondingly low (Fig. 1). Subsurface temperatures and groundwater conditions were strongly influenced by the presence of extended surface and subsurface ice. Glacial erosion in the ice-marginal zones was probably limited due to strongly reduced basal sliding and melt-water flow. More humid conditions with higher flow velocities, more basal sliding and stronger erosion by abrasion and melt-water effects must have prevailed during ice advance across the Swiss Plateau, and rapid down-wasting or even collapse (calving instability in lakes) is likely to have taken place during the retreat phase back into the Alpine valleys (Fig. 2). It appears plausible to assume that similar cycles were characteristic for past Ice Ages in general and could also be characteristic for future Ice Age conditions in northern Switzerland. Some key questions concerning effects of deep glacial erosion are formulated.

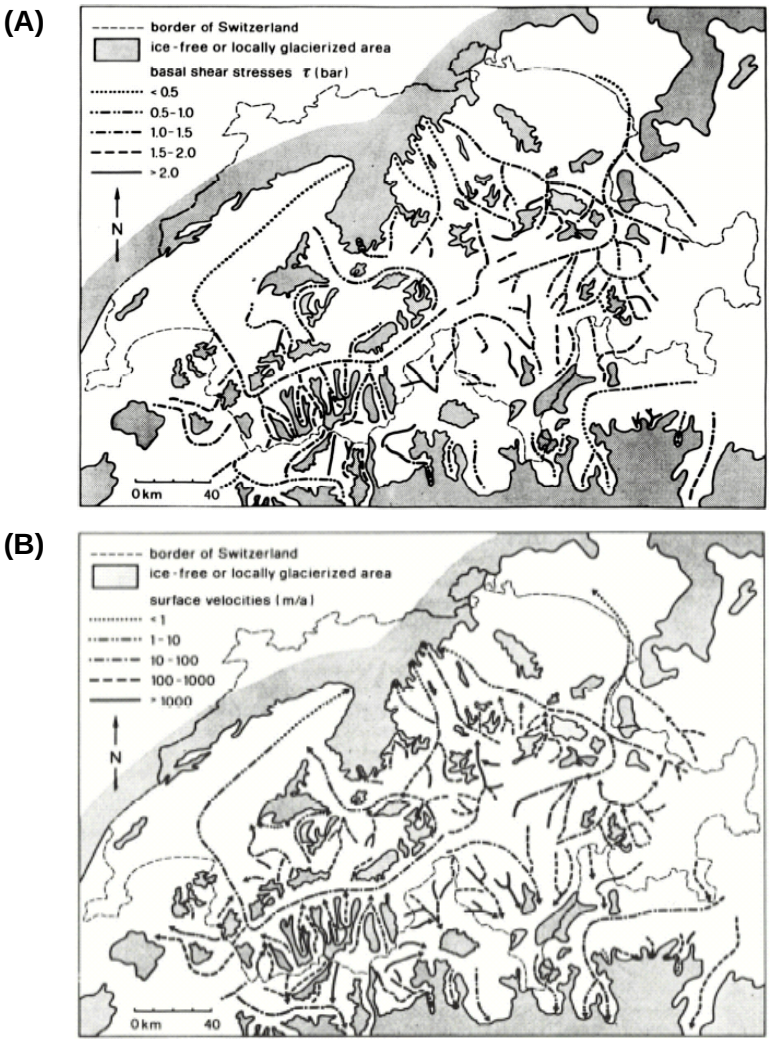

Fig. 1: Ice Age glaciers (18 ka BP) in and around the Swiss Alps. (A) Basal shear stresses. Values are averaged over distances greater than 10 times the local ice thickness. (B) Surface velocities. Values are averaged over distances of 10 km (Haeberli & Penz 1985).

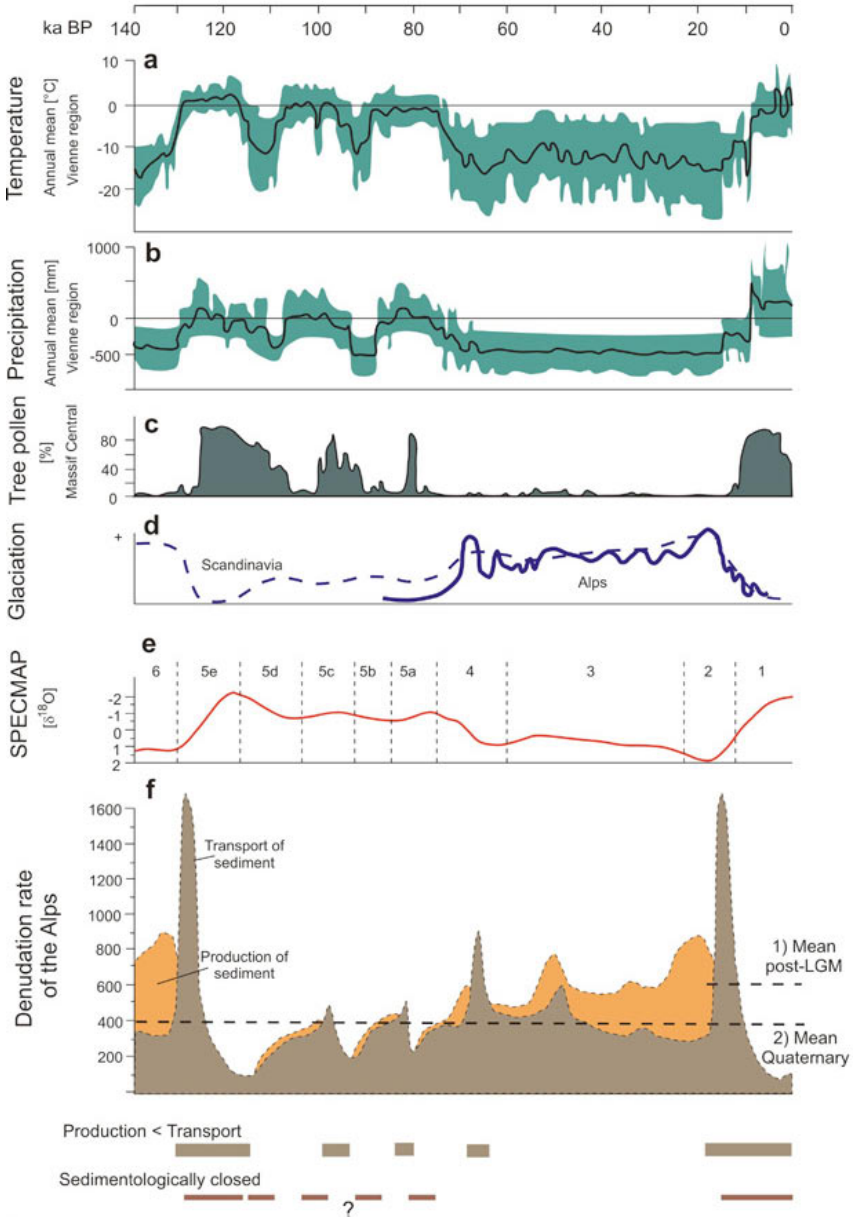

Fig. 2: Variations in climate (a–e) and denudation rates (f) in the Alps during the past 140 ka based on Quaternary and post-LGM sediment volumes (Hinderer 2001).

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Constraining the plausible amount of future glacial erosion

Bernard Hallet

The amount of erosion by the former Northern Hemisphere Ice Sheets (NHIS) has received considerable attention from the geological and glaciological community, but the subject has proven challenging because subglacial erosion is inherently difficult to study for a number of reasons: (i) where the beds of former ice sheets are exposed, past events are effaced with time as the surface erodes, a problem common to all eroding landscapes; (ii) where these processes are active at present, that is, at the bottom of ice-sheets, they are difficult to observe; and (iii) glacial erosion varies in time and space and depends on conditions at the bed, but bed conditions of former ice sheets can only be inferred from sparse, residual geologic information.

Investigating this subject requires the use of all available information about the NHIS that pertains to erosion, as well as studies of glacial erosion in other regions combined with theoretical considerations. My presentation (i) summarizes geological studies that use both traditional and new techniques in the area formerly occupied by the NHIS (e.g. Sugden 1976; Bell & Laine 1985; Hay et al. 1989; Lidmar-Bergström 1997); (ii) discusses results of empirical studies of glacial erosion on bedrock and sediment substrates in diverse settings (e.g. Jahns 1943; Carol 1947; Briner & Swanson 1998); it includes extreme cases of deep erosion by ice and by catastrophic glacial outburst floods; (iii) outlines theoretical considerations of glacial erosion (Fig. 1) and their application in a model of erosion by the NHIS (Hildes et al. 2004); (iv) considers a particular case example where considerable evidence exists for erosion by subglacial meltwater (e.g. Kor et al. 1991; Shaw 2002) and for the occurrence of a sediment cover over the bedrock, which impacts bedrock erosion; and (v) presents relevant results from the University of Toronto Glacial Systems Model (e.g. Tarasov & Peltier 2004, 2007), uses them to compute plausible ranges of total erosion for 100’000 years in southern Ontario, Canada (Fig. 2), and ends with a short discussion of the limitations of numerical models of erosion.

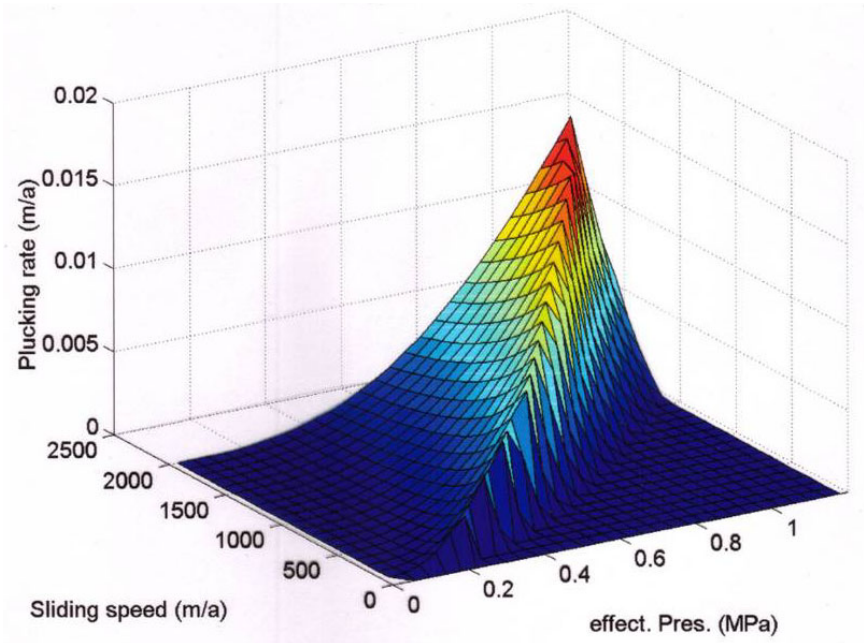

Fig. 1: Quarrying model results showing the plucking rate as a function of sliding speed and effective pressure.

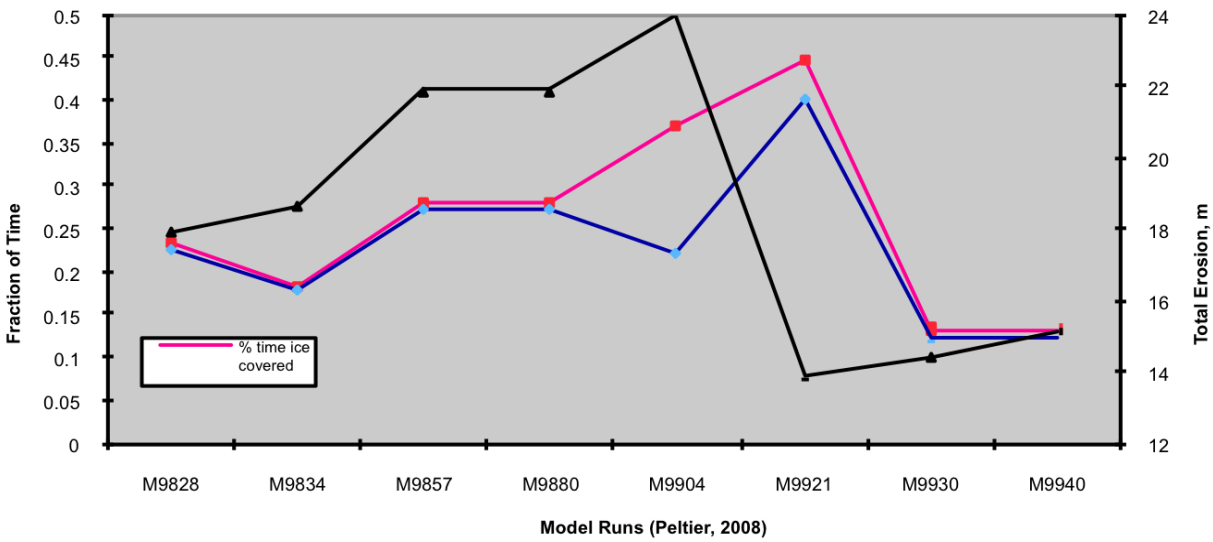

Fig. 2: Results of 8 University of Toronto Glacial Systems Model runs showing the fraction of time (left axis) the Bruce Peninsula is covered with ice (magenta) and is at the melting point (blue) as well as the cumulative erosion (right axis) in 120 ka (black).

Summing up, the state of understanding of erosion and other processes occurring at the base of ice sheets is far from complete, and the subglacial conditions that control erosion are likely to vary with time and space in complex ways, hence the magnitude of erosion over 100'000 years cannot be assessed with precision. Many lines of empirical evidence and theoretical results, however, point to a remarkably coherent conclusion that bedrock erosion on this time scale is likely to average between a few meters and a few tens of meters (Fig. 3). Much deeper erosion, however, can occur locally where erosion by water or ice would be greatly intensified, such as where the glacial setting and topography conspire to create conditions favourable for the recurrent formation of large ice dams that tend to fail and cause catastrophic floods (Fig. 4).

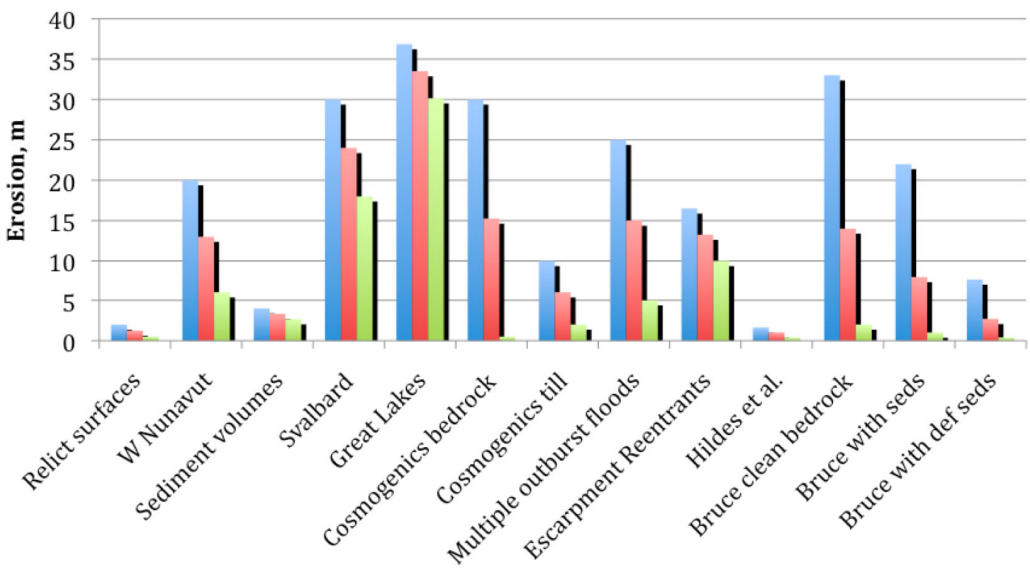

Fig. 3: Estimates of the amount of erosion in 100'000 years from different studies using a variety of approaches.

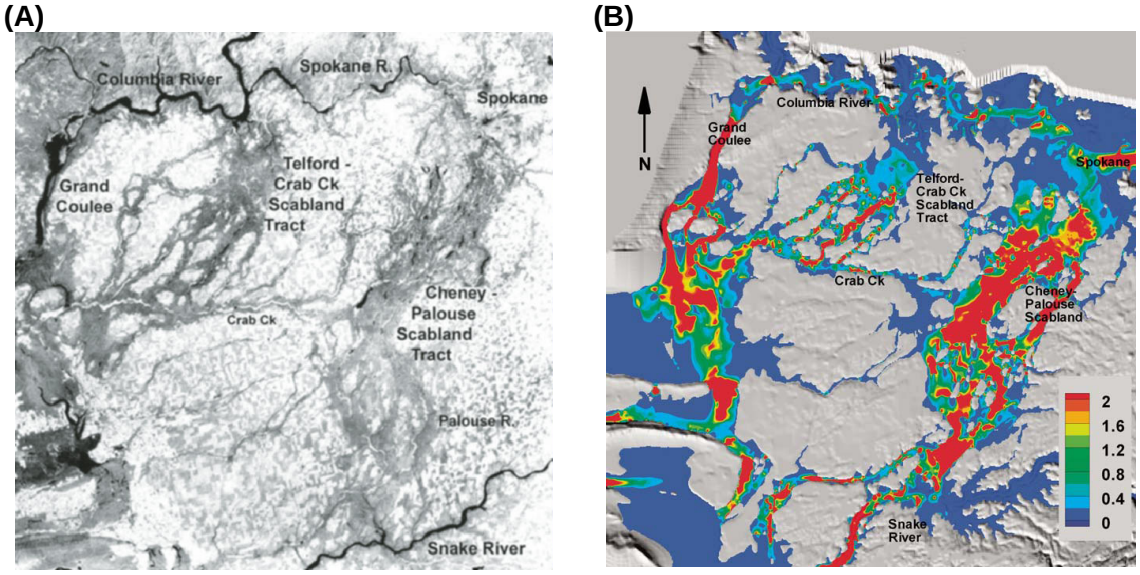

Fig. 4: (A) Landsat MSS satellite image of eastern Washington showing the Columbia and Snake Rivers, and the distribution of the Channeled Scablands across the arid prairie. The scablands show up as dark-gray scars on the landscape, as the soil has been stripped and the bedrock (basalt) eroded into characteristic channels of scabland morphology. (B) Flood power (truncated at 2 kW/m^2) during maximum simulated inundation of the Channeled Scablands 23 hours after rupture of the ice damming Glacial Lake Missoula. Distribution of modelled power per unit area (product of mean velocity and bed shear stress) approximates the pattern of flood scars in (A) (Denlinger & O'Connell 2010).

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Effects of glacier hydrology and sediment transport on erosion patterns

Frédéric Herman, Flavien Beaud, Pietro Sternai

Glaciers have a very strong impact on landscapes, carving U-shape valleys, creating big steps in the longitudinal profiles and large overdeepenings. Although their imprint is qualitatively well described, the mechanisms involved are still poorly understood. Until today most of glacial erosion models are able to reproduce glacial valley shapes to a first order but fail to explain steps and overdeepenings. Here we couple a 2D first order numerical ice model (Blatter 1985) to water pressure changes due to the evolution of the basal hydrology (Flowers & Clarke 2002a, b), that we apply to a valley profile. In turn, water pressure at the base lowers the basal shear stress which leads to an increase of basal sliding velocities and then erosion rates.

It appears that most of the basal water is located below the ELA, which causes velocities to increase in the ablation zone. It turns out from our modeling results that erosion is not only concentrated around and above the ELA but can also be large down to much lower altitudes. It is important to stress that using such an approach enables us to produce steep steps and some overdeepenings (Fig. 1). Comparing these results to actual profiles, we find that modeling basal hydrology is a key component in understanding glacial erosion.

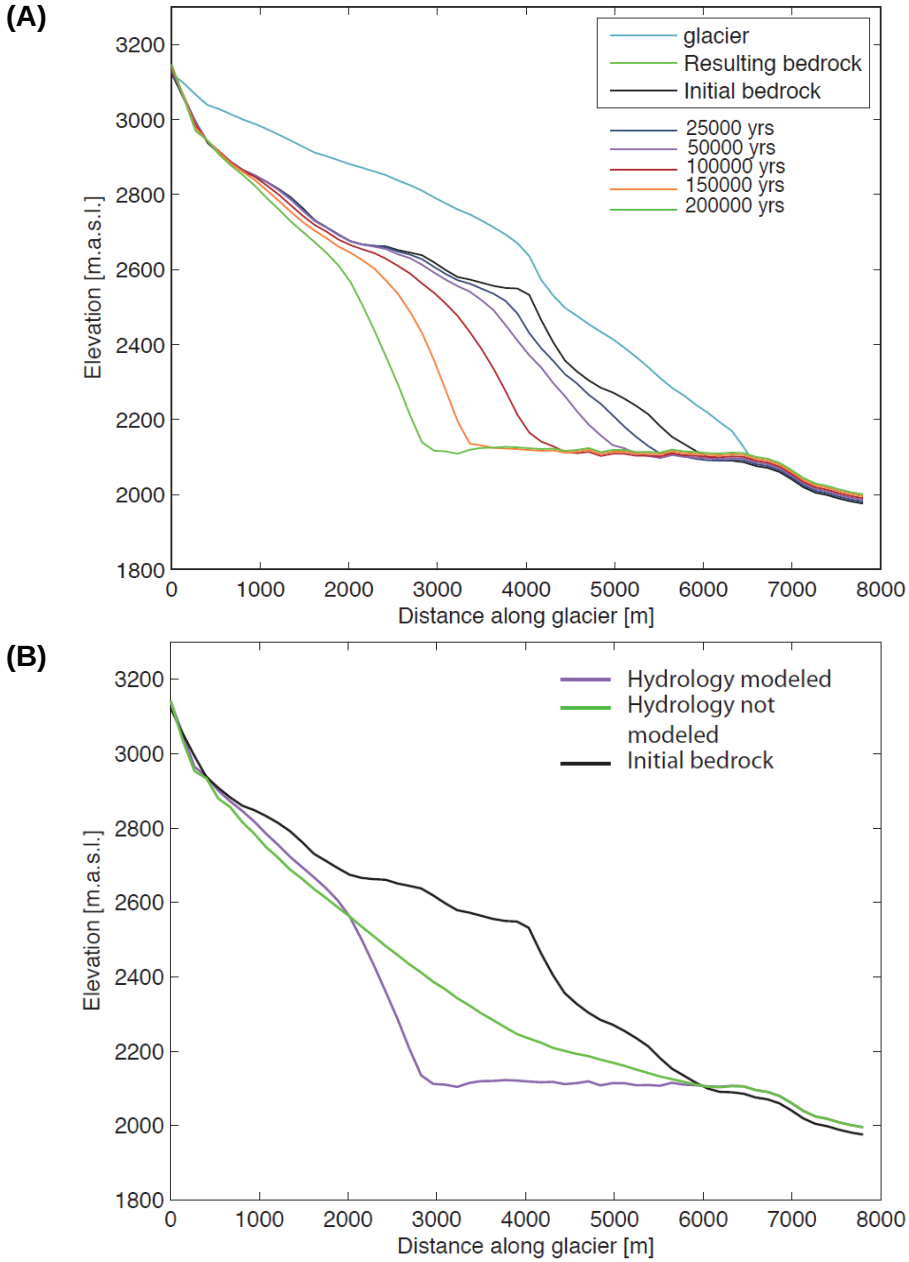

Fig. 1: Simulated long-valley profile evolution. Initial bedrock profile is shown as black lines. (A) Erosion of bedrock profile in 5 time intervals. (B) Influence of including basal hydrology on the bedrock profile after 200000 years.

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Modelling glacial erosion: heat, friction and water

Nick R.J. Hulton

This paper considers how the modelling of glacial erosion is sensitive to the way in which relationships between heating, friction and basal water production are represented in ice sheet erosion models. To the first order, ice sheet erosion rates are conceived to be dependent on the ability of ice sheets to slide decoupled from the bed. It is widely recognised that in the vast majority of cases, this process works efficiently when water, particularly pressurised water exists at the bed. This can come from penetration of surface water, or it can be produced (and maintained) at the bed if it reaches pressure melting temperature. Thus a first order prediction of where erosion occurs is determined by an accurate prediction of where the bed reaches pressure melting point and is able to produce excess heating for further melt at the bed.

Basic representations of temperature-sliding dependency in ice sheet models often rely on a simple switching behaviour such that sliding starts once the bed reaches pressure melting point (Fig. 1). Typically, a simple linear parameter is used to relate sliding rate to driving stress. The melting-point switch can result in the tendency for the ice sheet to produce localised temporal oscillations between sliding (warm) and non-sliding (cold) modes. This modelled behaviour is superimposed on further unstable behaviour associated with creep instability. Typically, at a given thickness and surface gradient, enough heat is produced at the bed for it to reach pressure melting point. Sliding starts, itself producing more frictional heat which allows the pressure-melt state to persist. The sliding only switches off once the ice has thinned or flattened sufficiently that the heat production reduces temperatures below pressure melting point. Normally at this point the local ice profile has significantly lowered and flattened compared to the profile prior to the onset of sliding.

The period and intensity of sliding is related to the sliding rate parameter in such a model. If ice can slide easily for a given driving stress, then rapid heat production ensures that basal melting is sustained and ice is drawn down rapidly from the ice sheet centre. The sliding will only switch off when the ice has thinned significantly, and it will subsequently take a long time to recover a sufficient profile to allow enough heat to build up at the bed to let sliding commence again. A simple modification is to scale the sliding parameter to meltwater production as a proxy for water pressure. This has the effect of removing or significantly damping down the tendency for oscillatory behaviour because sliding rates turn on more slowly. The degree of temporal variability can be controlled by the way in which water production is scaled to sliding. However, this misses a further important feedback. As meltwater is produced and pressurises, the effective coefficient of friction at the bed tends to reduce, and thus the frictional heating rate is consequently reduced. Such a feedback is introduced by explicitly modelling water pressure as a consequence of meltwater production and ice thickness distribution, and then relating water pressure to a basal friction coefficient. In doing so, once more, much more stable sliding patterns are represented in the ice sheet model, but the nature of these can be controlled by the way the basal friction is related to water pressure. In addition, the modification of basal friction has implications for erosion processes. Overall, we demonstrate that the prediction of sliding behaviour, and thus assumed erosion processes, are quite strongly controlled by the way heat production and sliding rates are characterised within ice sheet models. In particular it is clear that the explicit prediction of water pressure at the bed can be a key control on where sliding occurs, and the nature of the frictional contact at the bed. These have clear ramifications for the simulation of erosion rates.

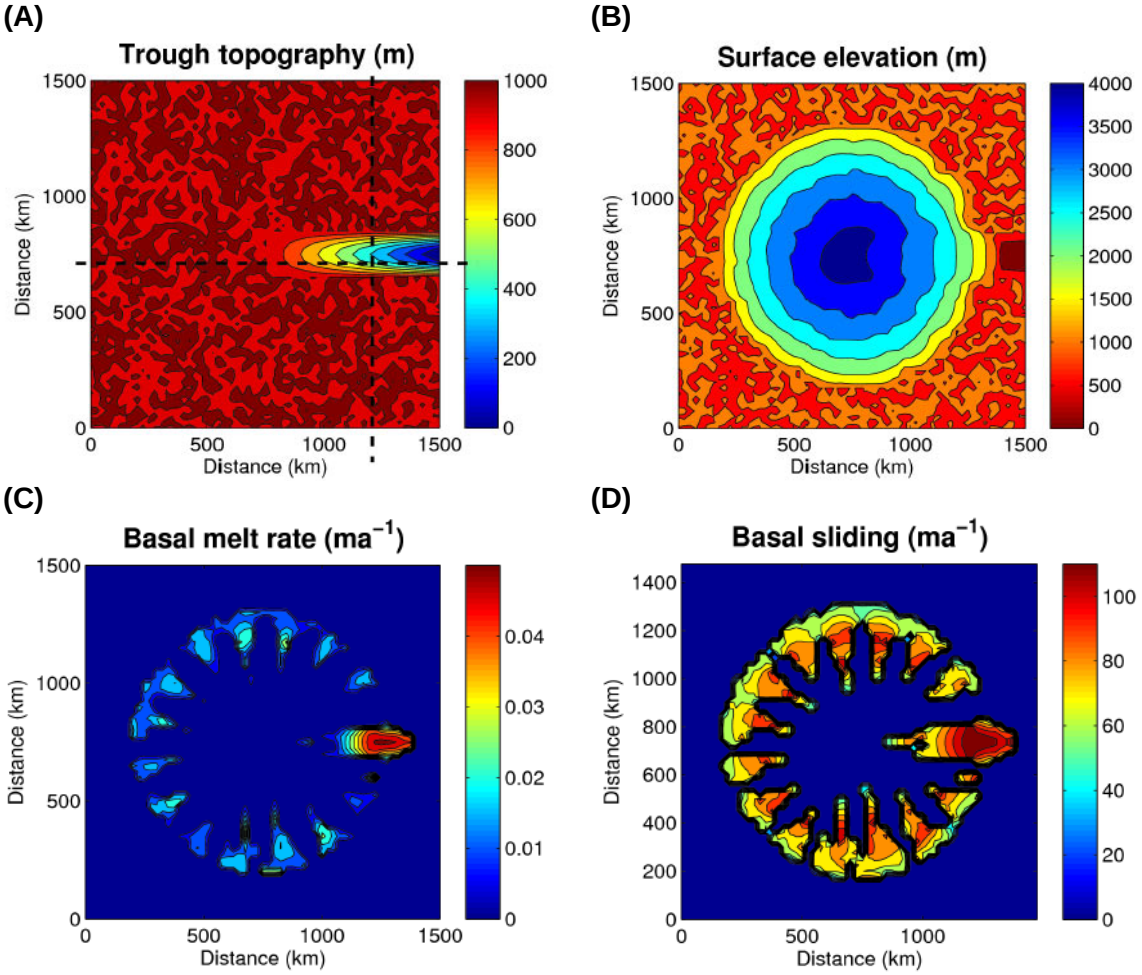

Fig. 1: Simulations using the "melting-point switch" (see text for details). (A) Basal topography. (B) Surface elevation of circular EISMINT-style ice sheet. (C) Basal melt rate when above the pressure melting point. (D) Basal sliding velocity.

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A model of quarrying based on adhesive wear and Weibull rock fracture

Neal R. Iverson

In models of glacial landscape evolution, erosion rate is usually assumed to equal the product of the glacier sliding speed (or ice discharge) and a coefficient that depends on the erodibility of the rock bed. Quarrying is widely thought to be the principal mechanism of glacial erosion, and observations of glaciated bedrock indicate that preglacial fractures control the orientation and geometry of quarried surfaces. However, existing quarrying theories do not include the spatial variability of bedrock strength that results from these fractures.

Strength heterogeneity of bedrock is included in a new model of quarrying grounded on the theory of adhesive wear (breakage of irregularities along sliding brittle surfaces) and a Weibull statistical description of bedrock strength. The model includes ice-bed separation and its effect on both stresses between ice and rock and the probability of exploiting a fracture in the bed. Results indicate that quarrying rate and its dependence on sliding speed and effective pressure are highly sensitive to bed-strength heterogeneity (Figs. 1 and 2). No single "erosion rule" of landscape evolution models, therefore, is expected to have widespread applicability because the form of the rule will depend on bedrock lithology and structure.

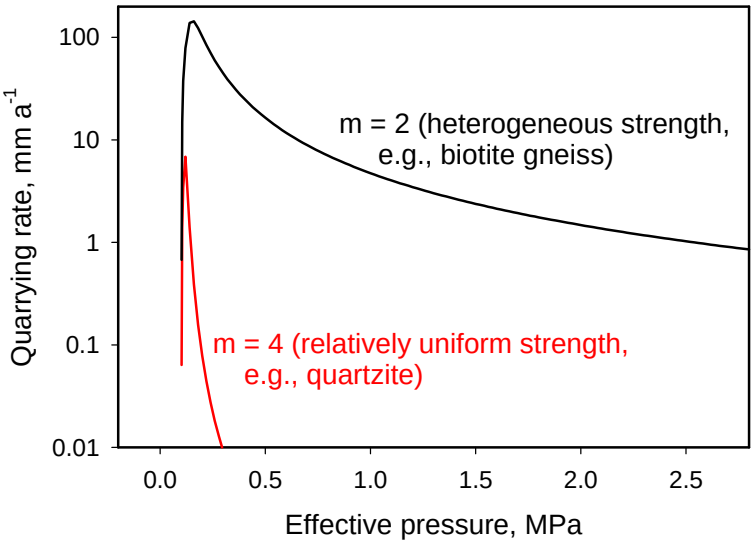

Fig. 1: Quarrying rate as a function of effective pressure for two different types of bed-strength heterogeneity (*m* denotes the Weibull coefficient of uniformity) calculated for a glacier sliding with a velocity of 20 m a^{-1} over a bed with uniform steps of height 1 m and 10 m apart.

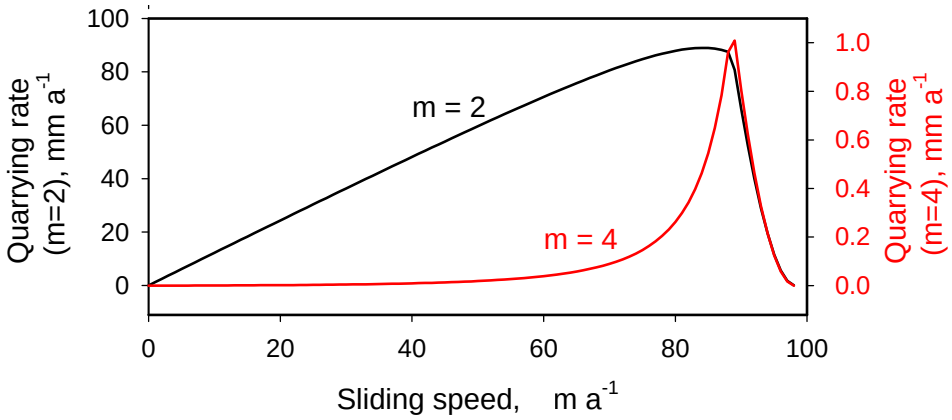

Fig. 2: Quarrying rate as a function of sliding speed for two different types of bed-strength heterogeneity (*m* denotes the Weibull coefficient of uniformity) calculated for a glacier sliding at an effective pressure of 0.5 MPa over a bed with uniform steps of height 1 m and 10 m apart.

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Modelling long-term glacial landscape evolution: feedbacks between ice, erosion and topography

Stewart S.R. Jamieson, Nick R.J. Hulton, Magnus Hagdorn, David Sugden

The aim is to understand how topography influences ice behaviour and erosion, and how the glacially evolving landscape then impacts upon ice mass behaviour. We use a 3-dimensional ice sheet model (Rutt et al. 2009) with a coupled hard bedrock erosion system (Jamieson et al. 2008). The model uses the shallow ice approximation and is fully thermodynamically coupled, allowing thermal feedbacks caused by erosion and basal friction to be included. Model resolutions are typically 5 to 40 km depending on the domain size, and are therefore are best suited for understanding ice sheet scale systems (Jamieson & Sudgen 2008, Jamieson et al. 2010). We assume erosion requires water at the bed via basal melting and that the relationship between basal friction and basal melt-rate is a smooth curve whereby friction decreases less and less rapidly until a maximum is reached. Our approach is akin to assuming the bed has a variable permeability, and means that the basal system can neither be overpressurised (which would lead to a frictionless bed) nor pressurised too rapidly (which due to the difficulties in incorporating a full treatment of subglacial hydrology would lead to unrealistically rapid transitions in modelled basal velocities). Modelled erosion is a function of basal ice velocity (and therefore basal friction and basal melt-rate), ice thickness and erodibility, and therefore the importance of basal water for landscape evolution is incorporated in a simple manner. Modelled rates of erosion scale with time due to the erodibility parameter so timescales for modelled glacial transformation are uncertain.

We model the development of a circular ice sheet upon a range of synthetic landscapes including: (i) a landscape with existing radiating valleys; (ii) a topography with valleys oriented transverse to predicted ice flow; and (iii) a modelled 'fluvial' landscape. In doing so we ask: *does existing topography determine the distribution of overdeepenings?* We further ask: *can new valleys be cut in patterns that do not follow the local/regional grain of a landscape?*

With a radial topography, we unsurprisingly find that selective erosion naturally exploits existing valley topography. By varying the relief in the system we see that even small valleys will eventually generate preferentially focussed flow and will develop overdeepenings. With a transverse topography however, it is possible to cut new radial valleys through topographic barriers that have significant relief (Fig. 1). Where ice is sufficiently thick and overrides the landscape then ice surface slope is the 1st order control on the alignment and development of overdeepenings, allowing new glacial troughs to be incised. We suggest that if ice is thinner or the relief is stronger, then ice surface slope becomes increasingly less important in controlling valley incision. It is not clear how important the relative angle between existing valley orientation and ice surface slope is. Experiments are planned to systematically test how ice flow and erosion are perturbed as existing valley alignment is varied in relation to ice surface slope and ice thickness. However, if one considers the arrangement of large-scale fluvial drainage systems in mountain regions and the results of the radial topography experiment above then it seems likely that glaciers will follow existing drainage systems. Thus it becomes difficult to envisage new overdeepenings being cut in a pattern that is significantly against the grain of existing topography unless an ice mass grows in a location that is offset from the long-term average position of ice masses or unless another valley is cut by other means (e.g. rivers, faulting etc.) between glaciations. Understanding variations in local and regional accumulation patterns over successive glacial cycles, and the resulting positions scales of ice masses is therefore critical.

With an existing fluvial topography, we see the importance of existing topography in developing the erosional regime. Figure 2 illustrates how a system can undergo the change from wide area erosion (areal scour) towards selective erosion to produce valley overdeepenings. The implication is that given enough time, and a lack of interglacial 'resetting', topographies that promote convergent ice flow will eventually generate overdeepenings. The basal thermal regime of the ice sheet changes in response to overdeepening development – as ice gets thicker in the troughs, the base gets warmer and erosion is more likely to be re-enforced. However, because we limit the maximum influence of basal-melt rate some basal friction is always maintained so the ice never becomes fully-decoupled from its bed. Under such conditions of intensely focussed erosion, and over long timescales, an ice sheet can become less extensive as a result of mass balance feedbacks. Such a feedback has been hypothesised to have been responsible for reductions in ice sheet extent over successive glacial cycles in Patagonia (Clapperton 1990).

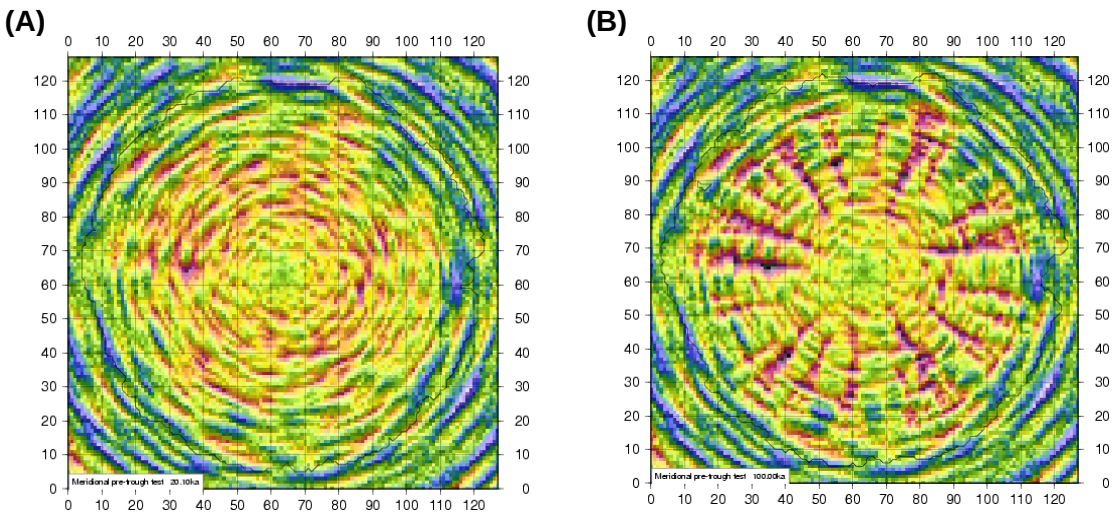

Fig: 1: (A) Initial topography depressed under load of circular EISMINT-style ice sheet. (B) 'Glaciated' topography displaying radially overdeepened troughs that align with ice surface slope and cross-cut existing topography. Colour shows elevation: red=low, blue=high.

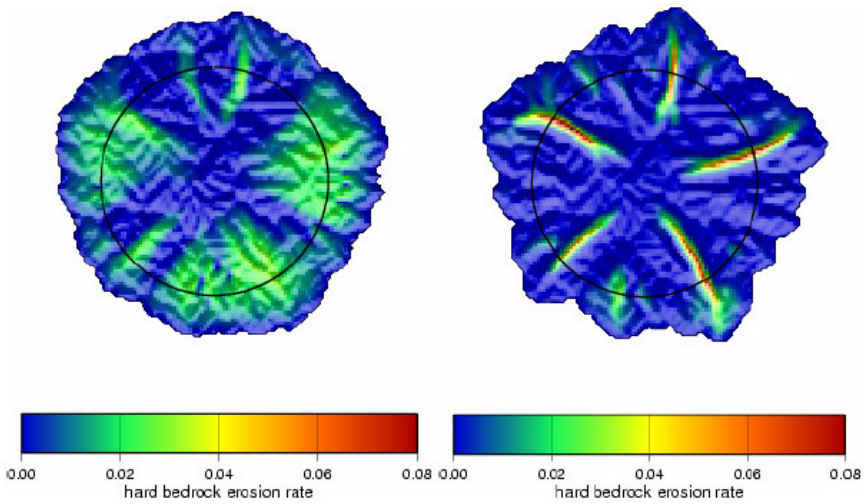

Fig. 2: Transformation of glacial erosion regime from areal scouring (left) to selective erosion (right) as a 'fluvial' landscape is subjected to glaciations under a circular ice sheet. Note erosion rate can be scaled with time. Black circles show the initial position of the ELA.

We conclude that existing topography is a key influence in determining the pattern of glacial overdeepenings and that ice will exploit the easiest pathway to the ablation area. The relief of the pathway need not be large to determine ice flow directions and the erosion regime will eventually become increasingly focussed as topographic steering is enhanced under increasing trough depth. However, if an ice mass is sufficiently thick, then ice surface slope can override the importance of existing topography and new troughs can be cut. To do this, a change in the 'average' ice mass scale and/or position would be required (e.g. via a shift in local or regional climate patterns), or a perturbation must be introduced at the bed (e.g. by interglacial or tectonic processes). Once overdeepening becomes established, frictional heating at the ice base causes the pattern of erosion to become entrenched and increasingly selective. This may cause an ice sheet to reduce its extents.

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Empirical measurements of basin-averaged glacier erosion rates

Michele N. Koppes

A new global compilation of erosion rates refutes the conventional contention that glaciers are more effective erosional agents than rivers. The range of basin-averaged erosion rates for both fluvial and glacial basins at catchment scales of 1-10^4 km^2 span 0.1 to > 10 mm a^{-1}, and tend to increase with (i) basin size and (ii) the transition from polar to temperate ice regimes (Fig. 1). The highest erosion rates, averaging up to 40 mm a^{-1} for the past 50 years, have been measured from rapidly retreating tidewater glaciers in Alaska and Patagonia (Fig. 2). A comparison of glacial erosion rates over time scales ranging from 10^1 to 10^6 years shows that basin-averaged erosion rates tend to decrease by a factor of 4-5 when averaged over one glacial advance-retreat cycle, and by at least one order of magnitude over million year time scales in tectonically active orogens, with up to a two order of magnitude decrease in tectonically quiescent orogens. Rates of river erosion, on the other hand, exhibit no apparent time-scale dependence (Fig. 3). That the highest rates of glacial erosion, far exceeding 10 mm a^{-1}, occur during rapid retreat suggests that contemporary measures of glacial erosion reflect an anomalously dynamic and erosive period since the end of the Little Ice Age.

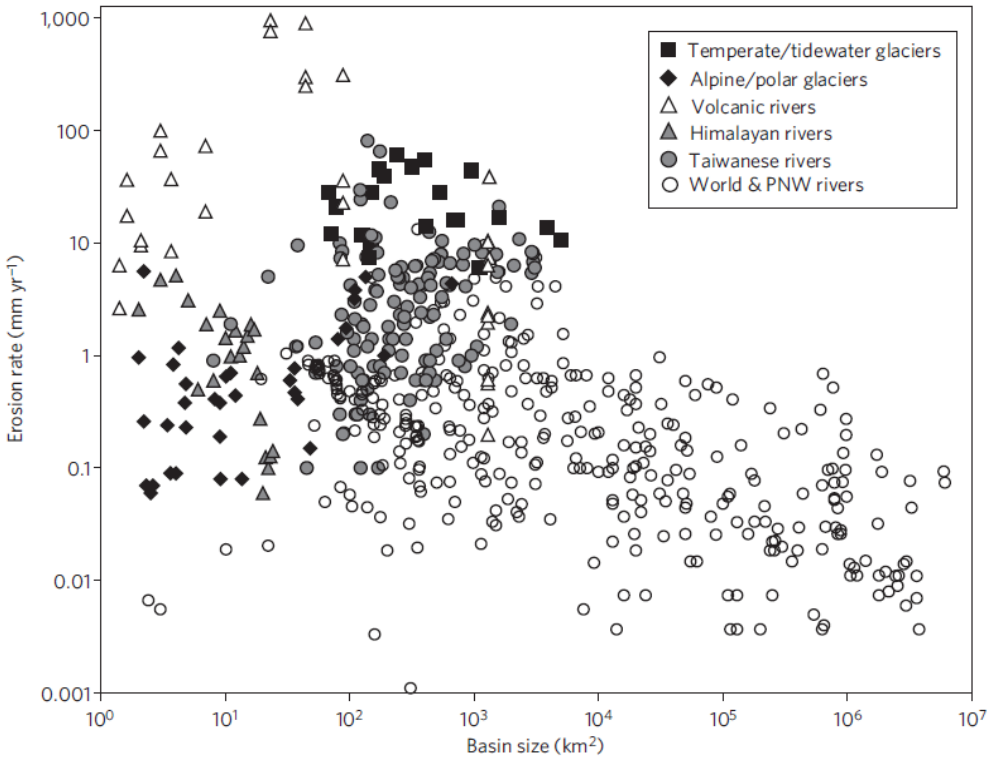

Fig. 1: Comparison of glacial, fluvial and composite landscape erosion rates versus contributing basin area, as measured by sediment yield data collected over 1-20 years. Fluvial basins are represented by circles and triangles: world rivers and basins in the Pacific Northwest (PNW) are open circles; fluvial catchments in tectonically active orogens are grey circles and triangles and volcanic rivers are open triangles. Glaciated basins are indicated by black squares and diamonds (Koppes & Montgomery 2009).

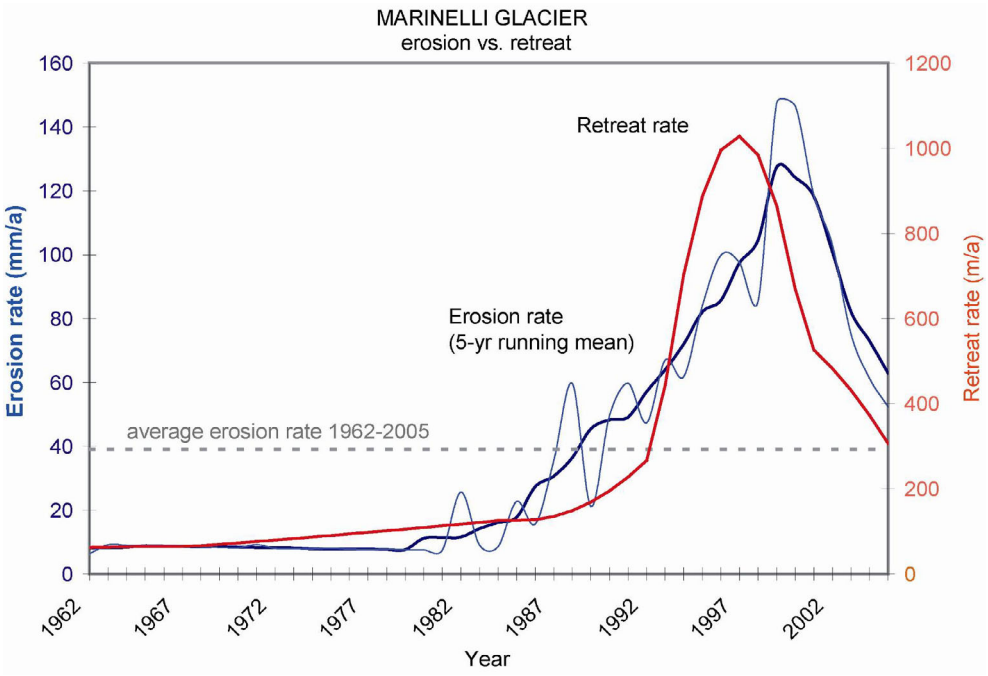

Fig. 2: Time-series comparison of erosion rate and retreat rate for Glaciar Marinelli since 1962. The contemporary erosion rate averages 39-12 mm a^{-1}. Extrapolating the erosion rate to times when the glacier is effectively stable, on average neither advancing nor retreating, the long-term erosion rate is ~ 10 mm a^{-1} (Koppes et al. 2009).

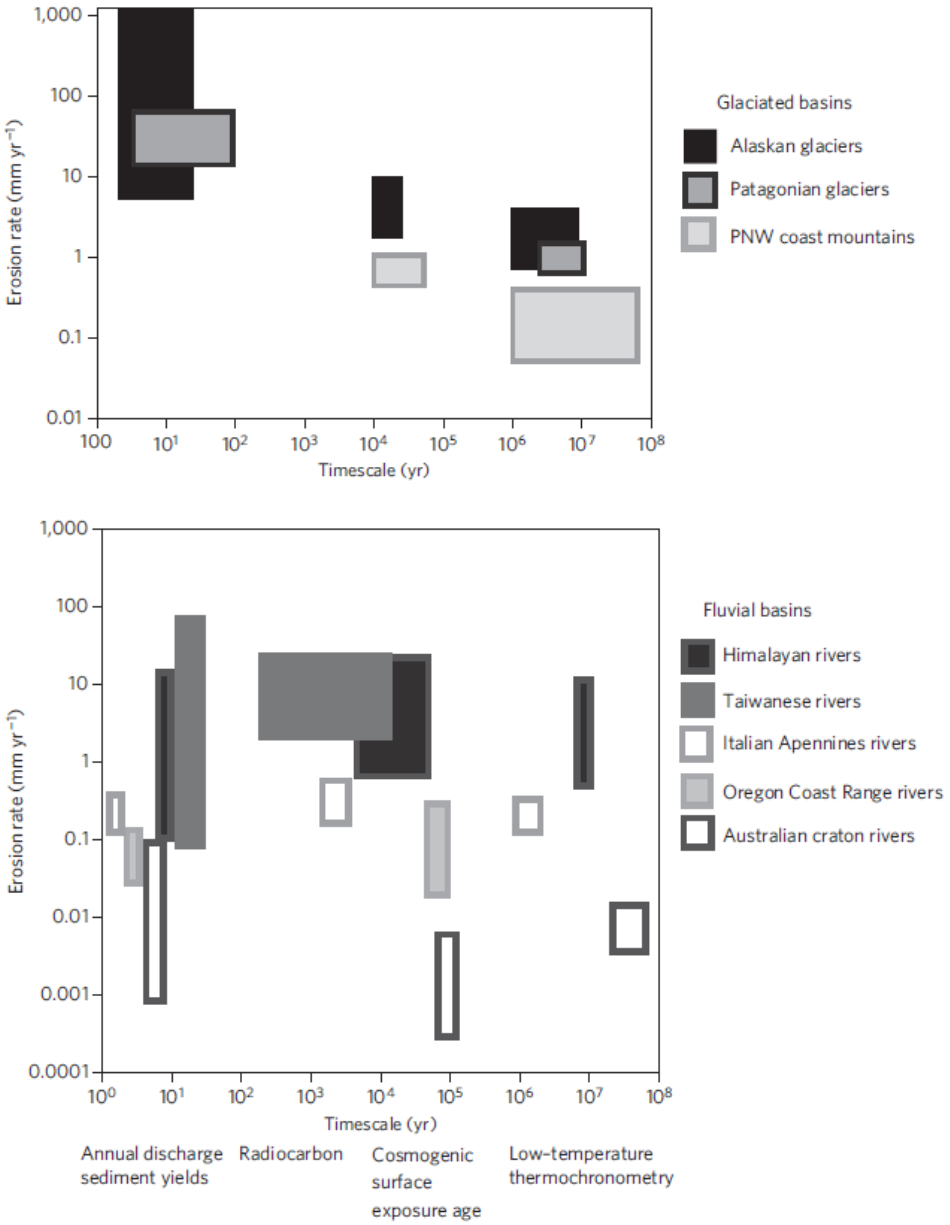

Fig. 3: Comparison of short-term and long-term erosion rates from glaciated and fluvial basins. Boxes represent ranges of erosion rates, including errors in estimates (height) and timescale of measurement (width). Top panel: Erosion rates measured from the same or proximal glaciated basins in Alaska, Patagonia and the coast mountains of Washington State in the Pacific Northwest (PNW). Bottom panel: Erosion rates measured from the same or proximal fluvial basins in orogens ranging from most tectonically active to most passive: the central Himalaya, Taiwan thrust belt, Italian Apennines, the Oregon Coast Range and the Australian craton (Koppes & Montgomery 2009).

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Simple modeling of quarrying and abrasion along the glacier longitudinal profile

Kelly R. MacGregor

Development of rock relief by glacial erosion in alpine landscapes is a key question in understanding the controls on mountain range topography. Removal of rock and generation of relief by glaciers at the high elevations in mountain ranges should shape atmospheric patterns of air-flow and the distribution of precipitation in alpine landscapes. We address relief production at high-elevation mountain crests using a numerical model incorporating glacier dynamics and subglacial erosion, as well as relevant hillslope processes. We include blowing snow from a valley-bounding plateau, and avalanching onto the glacier accumulation zone (Fig. 1). We use temperature-dependent frost cracking as a proxy for headwall backwearing. Our parameterization of subglacial erosion produces about 500 m of erosion over the 400 ka of each numerical simulation. In both steady and varying climate scenarios, valley floors flatten as headwalls increase in height and retreat headward over time. Headwall steepening occurs when quarrying is explicitly incorporated into the simulations (Figs. 2 and 3); quarrying focuses subglacial erosion rates locally, and when elastic flexural uplift is imposed in response to mass unloading. Small overdeepenings form at the base of the headwall in steady climate simulations, when the glacier is small at the end of the model run. While the final profiles are relatively insensitive to the erosion rule used, quarrying is most effective near the head of the glacier, whereas abrasion rates reflect the instantaneous pattern of integrated ice discharge (Fig. 3). In simulations in which cirques are formed, the equilibrium line altitude (ELA) is hundreds of meters above the cirque floor; however, the time-averaged horizontal position of the equilibrium line (EL) corresponds well with the downvalley cirque position.

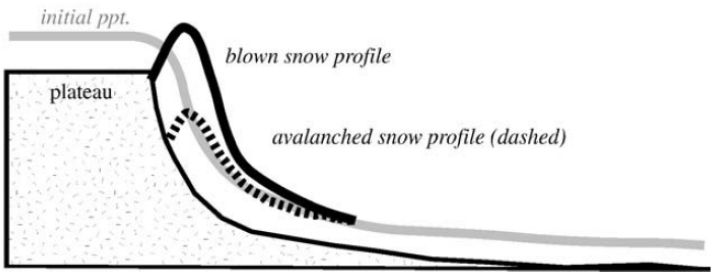

Fig. 1: Schematic diagram of the effect of blowing and avalanching snow on the pattern of precipitation. Blowing snow off the plateau is distributed in an exponentially decaying pattern on the valley below. Headwall slopes > 25º are cleared of snow via avalanches. (MacGregor et al. 2009).

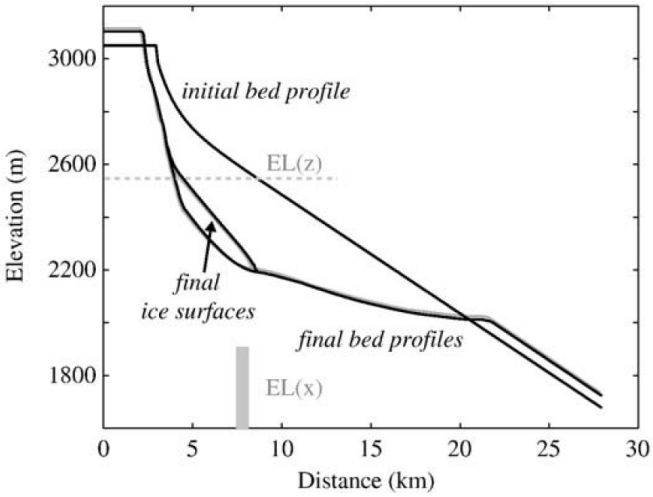

Fig. 2: Steady-temperature simulations using an abrasion-only rule showing the inital bed profile and final bed and ice surface profiles (MacGregor et al. 2009).

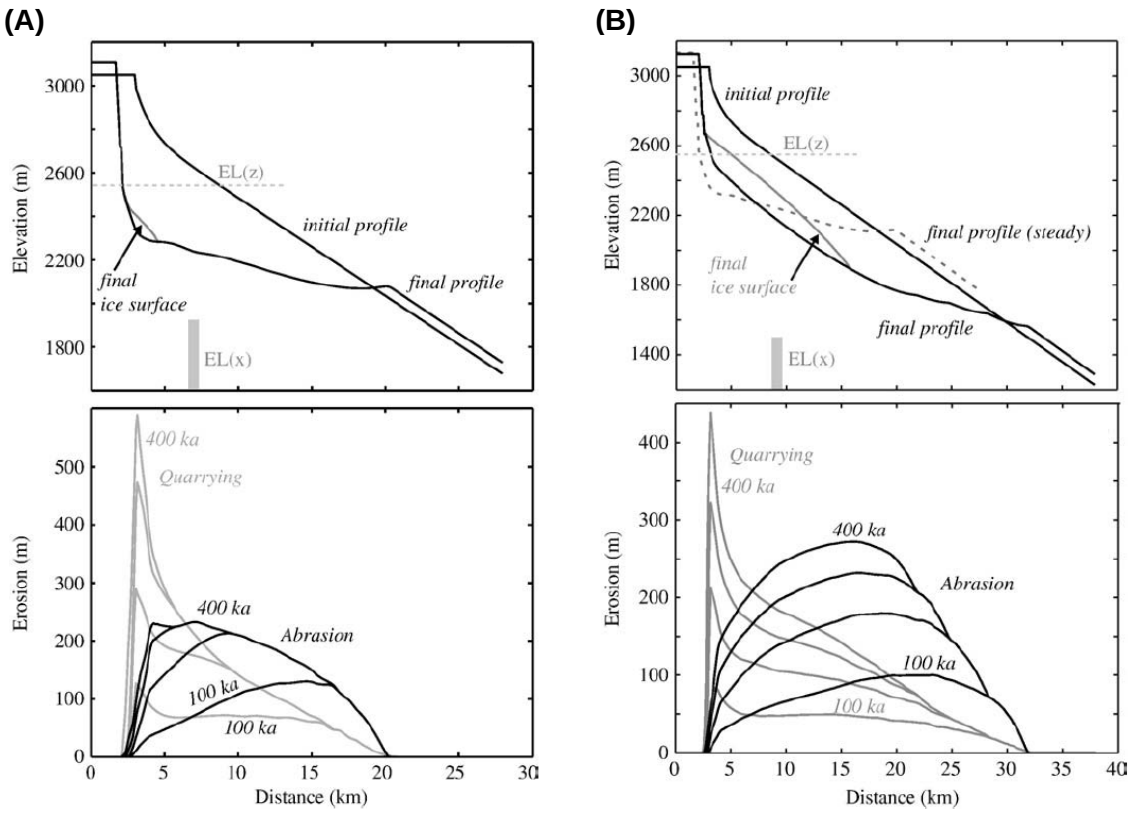

Fig. 3: Simulations using quarrying and abrasion rules with (A) a steady temperature and (B) variable temperatures. Top panels: Initial and final bed profiles (black lines) and final ice surface (grey). The headwall steepened and lengthened. However, during the variable-temperature simulation, less erosion occured at the base of the headwall, but greater erosion further downvalley. Bottom panels: The cumulative patterns of abrasion (black) and quarrying (grey). Quarrying is most important near the valley headwall and abrasion follows the spatial and temporal pattern of ice thickness (MacGregor et al. 2009).

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Coupling deformable sediment models with ice sheet models

David Pollard, Robert M. DeConto

A model of subglacial deformable sediment is added to a dynamical ice sheet model. The sediment model includes bulk transport under ice assuming a weakly non-linear till rheology, generation of new till by glacial erosion on exposed bedrock, and subaerial fluvial transport of previously disturbed sediment. Results are presented for two applications: (i) continental East Antarctica during first major Cenozoic ice growth, and (ii) idealized flowline simulations for marine ice over a continental shelf. All results are preliminary, and are intended only to demonstrate the feasibility of running coupled ice sheet-sediment transport models over long periods.

In the first application (Fig. 1), a terrestrial (Shallow-Ice-Approximation) ice sheet model over continental East Antarctica is driven through several million years by Global Climate Model look-up climates, with prescribed orbital variations and a gradual decline of atmospheric CO_2, simulating major Antarctic ice growth at the Eocene-Oligocene boundary ~ 34 Ma. At first, individual ice caps form on major mountain ranges and fluctuate on orbital periods, and later coalesce and grow into a continental ice sheet as CO_2 declines further. Sediment is initialized as a uniform continent-wide 50-m layer of regolith. Within ~ 1 million years, the early ice transports most sediment to the coast, funneled topographically mainly into several major outlet basins. In central regions, sediment transport is delayed as basal temperatures below the early ice remain below freezing, and reach the melt point only later when full continental ice cover is attained.

In the second application (Fig. 2), the ice model is extended to include ice shelf and stream flow, allowing grounding lines to migrate across submerged continental shelves. In the sediment model, glacial erosion of bedrock is allowed to lower the equilibrium bedrock topography. The coupled model is run over 10 million years with an initially linear bedrock slope, no initial sediment, and prescribed simple mass balance forcing with drastic short-lived warmings once every 1 million years that cause complete ice retreat. Sediment is generated by glacial erosion on inland bedrock, which is transported by subglacial deformation onto the submerged continental shelf. Early in the simulation, grounding lines are relatively shallow, which allows marine ice to extend further and deposit sediment packages on the distal outer shelf. Later, erosion of bedrock deepens the bathymetry, limiting the extent of marine ice and producing sediment deposition on the proximal inner shelf. The final marine sediment strata and reverse-slope shelf bathymetry are crudely reminiscent of observed Cenozoic marine sediments in West Antarctica. However, the basic form of the final sediment distribution changes drastically as uncertain sediment rheological parameters are varied, suggesting that the results depend strongly on the model of sediment transport used.

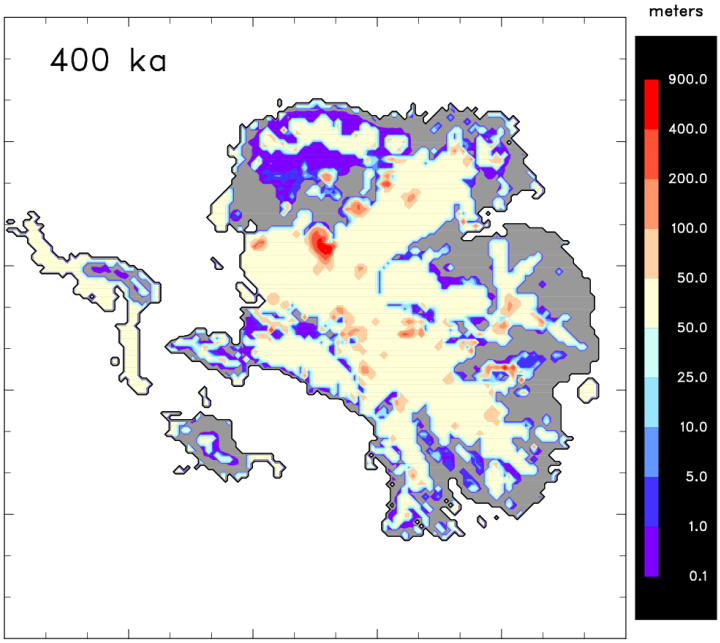

Fig. 1: Antarctic sediment thickness after 400 ka of a coupled sediment-ice sheet run. 50 m initial sediment thickness is shown as yellow, no sediment as grey. (Pollard & DeConto 2003).

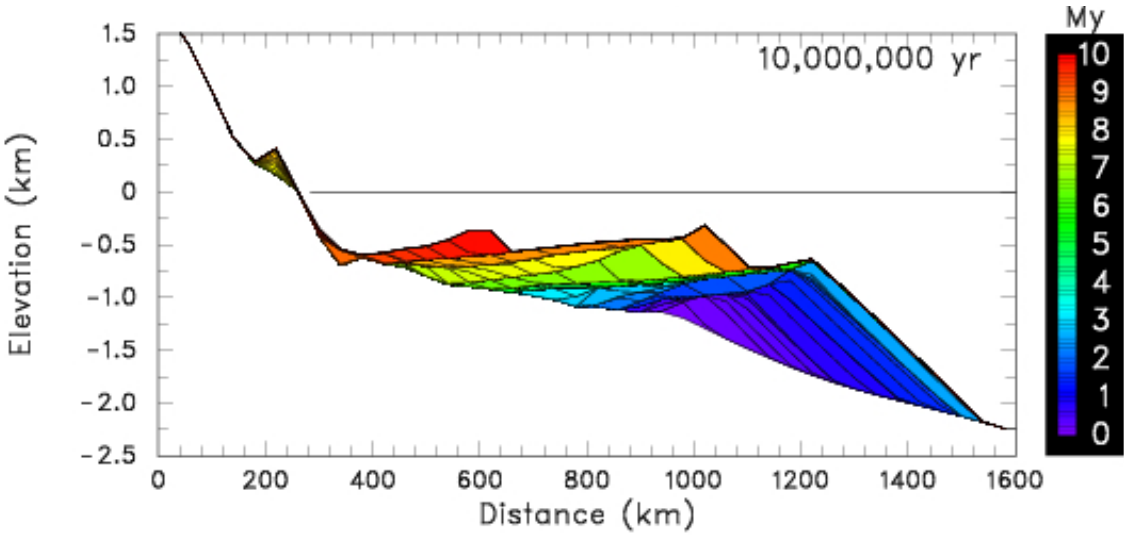

Fig. 2: Sediment strata at end of run, showing time of deposition from start of run. Horizontal line shows sea level at 0 m. (Pollard & DeConto 2007).

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Quaternary glaciation history of the Alps

Frank Preusser, Jürgen M. Reitner, Christian Schlüchter

One of the most important questions in Quaternary research in the Alpine realm is when the overdeepening of glacial valleys and basins occurred. To answer this question it is necessary to firstly understand the glaciation history of the area, i.e. when and where glaciers advanced from the Alps into the foreland. In this context, the Swiss Alpine Foreland represents the most complex geological record available. The earliest evidence for Quaternary glaciation is given by the multiphase gravels intercalated by till and overbank deposits ('Deckenschotter'). Mammal remains place the oldest of these deposits into Mammalia Neogene Zone 17 (2.5-1.8 Ma). The presence of till within the sedimentary sequences implies glaciations of the Swiss Alpine Foreland during the Early Pleistocene. Important differences in the base level of the gravel deposits allow distinguishing two sub-units ('Höhere Deckenschotter', 'Tiefere Deckenschotter'), separated by a period of substantial incision. Each of the sub-units contains evidence for at least two but probably even four individual glaciations. The Early Pleistocene is separated from Middle Pleistocene deposition by a time of important erosion, probably related to tectonic movements and/or re-direction of the Alpine Rhine. The Middle-Late Pleistocene comprises four or five glaciations, the oldest of which may represent the most extensive glaciation of the Swiss Alpine Foreland. The age of this glaciation is not well constrained but is believed to be older than Holsteinian (320 or 420 ka). First dating evidence implies another glaciation more extensive than the Last Glaciation Maximum at around 150 ka ago. The last glacial cycle probably comprises two or even three independent glacial advances dated to about 105 ka, 65 ka, and 25 ka.

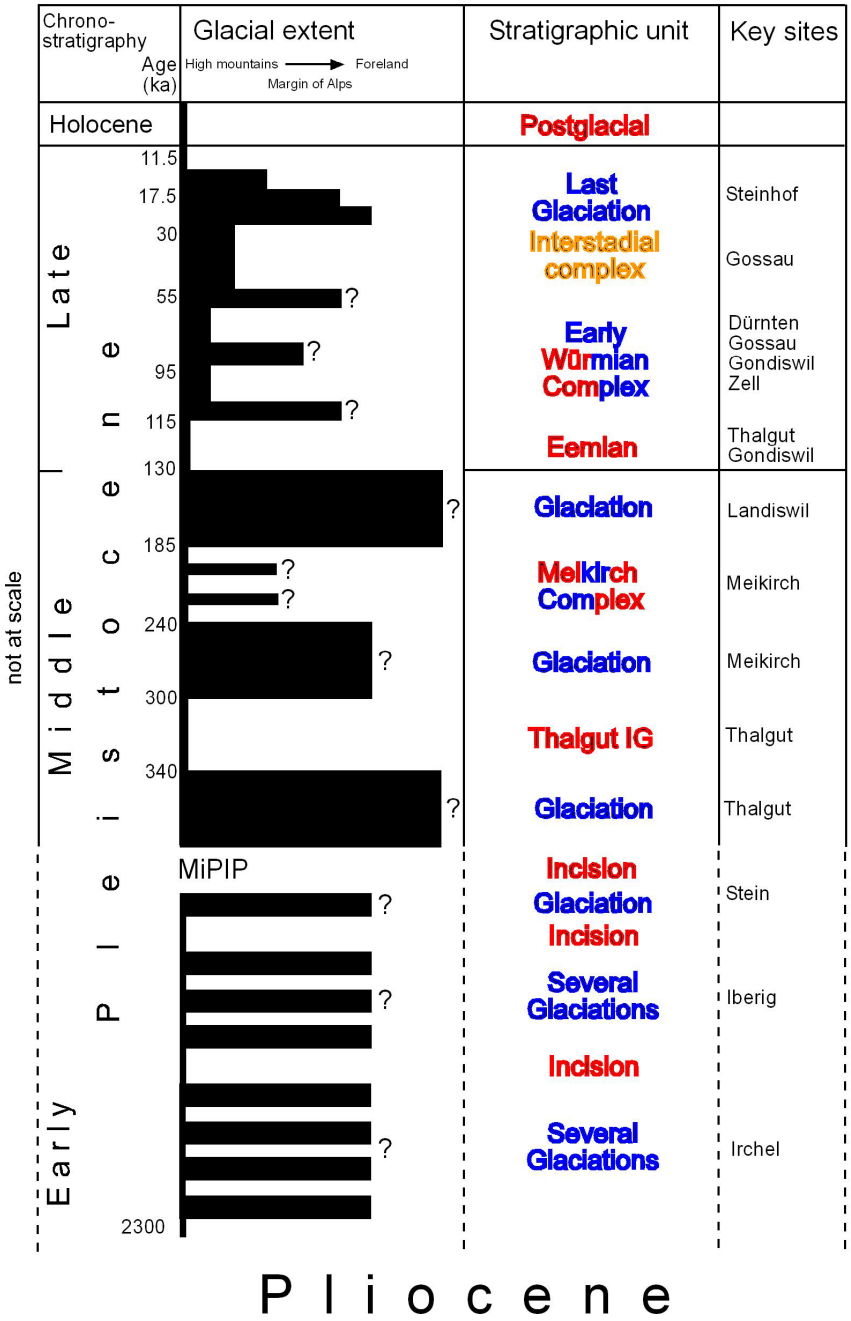

Fig.1: Schematic overview of the glaciation history of the Swiss Alpine Foreland.

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Background and past research activities with respect to deep glacial erosion in north-eastern Switzerland

Michael Schnellmann, Andreas Gautschi

The siting regions under consideration for a high-level waste repository are located in north-eastern Switzerland, where the host rock (Opalinus Clay, a Middle Jurassic claystone formation) dips gently to the south-east and occurs at depths between 400 and 1000 m below surface. Due to the time span to be considered for safety analyses (1 Ma), erosion was an important aspect for the selection of the three siting regions under investigation. The minimum bedrock overburden generally required for all sites is 400 m to account for the effects of uplift and (fluvial) incision. The Alps, with uplift and erosion rates > 0.5 mm/a, are excluded for a high-level waste repository. In north-eastern Switzerland, fluvial incision rates have been an average of < 0.2 mm/a over the last 2 Ma. This is evidenced by a geomorphological marker horizon ('höhere Deckenschotter') dated to 1.8-2.5 Ma.

The three siting regions proposed for further investigations are all located within the limits of the most extensive Pleistocene glaciation (Fig. 1). During the Last Glacial Maximum (LGM), the easternmost siting region was fully ice-covered, the central siting region was partly covered by ice and the westernmost siting region was ice-free. As an effect of glacial erosion, over-deepened valleys occur within the central and eastern siting regions (Figs. 2 and 3). The rock overburden is higher in the eastern and central siting regions (> 600 m over large areas) compared to the western siting region (mainly 400-550 m). The upper part of the rock overburden consists principally of Tertiary sandstones, siltstones and marls (Molasse), while the lower part includes 100-200 m thick Jurassic limestones.

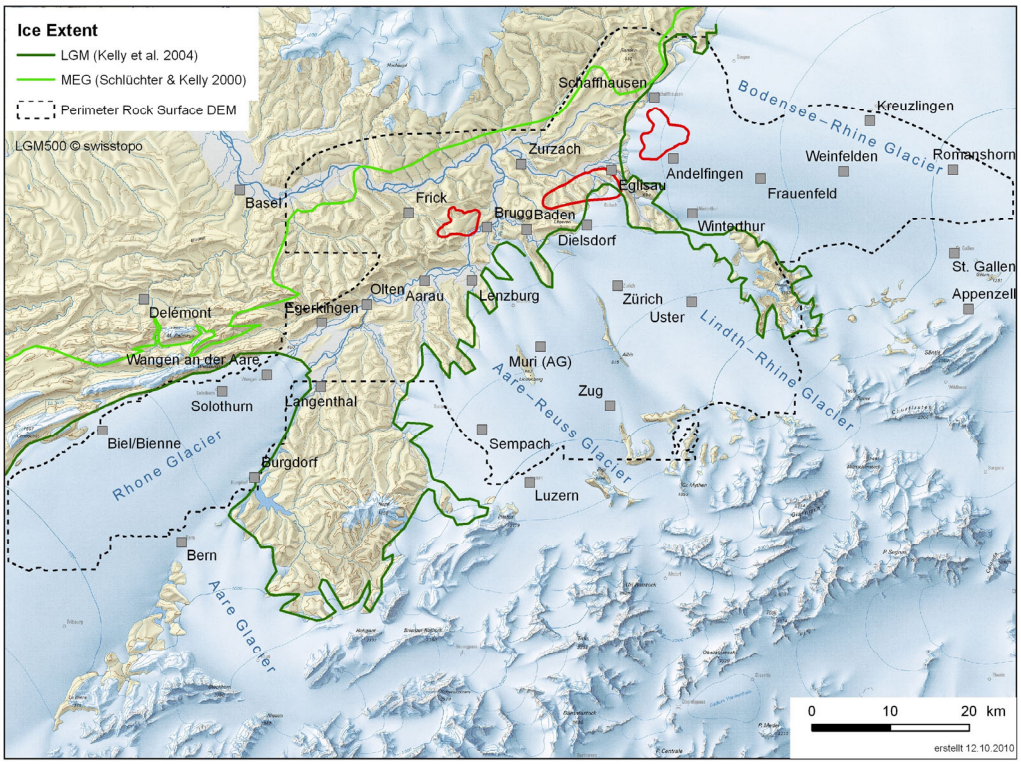

Fig. 1: Maximum ice extent in the Swiss Alpine Foreland during the Last Glacial Maximum (LGM; Schlüchter et al. 2009) and ice extent during the Most Extensive Glaciation (MEG; Schlüchter & Kelly 2000). Red polygons mark the proposed siting regions for a high-level waste repository.

Over recent years, a digital elevation model (DEM) of the top bedrock has been compiled for the Swiss Alpine Foreland starting from existing isomaps and taking into account thousands of boreholes (Jordan 2008, Jordan 2010; Figs. 2 and 3). The following observations seem noteworthy with regard to glacial erosion processes:

- The valleys in central Switzerland (related to the Reuss and Aare glaciers) are oriented parallel to each other and perpendicular to the Alpine strike (Figs. 1 and 2). In contrast, in the west (Rhone glacier) and the east (Bodensee-Rhine glacier) there is evidence for a radial flow component away from the ice lobes (Figs. 1 and 2).
- Most buried valleys are located below recent valleys, with few exceptions (Fig. 2).
- These exceptions, and the infill of the related valleys, document examples of changes in ice flow and related erosion over time. Analyses of glacial pebbles and heavy minerals also indicate changes in ice flow directions.
- Most of the major overdeepenings in the external Alpine Foreland occur in the Molasse substratum and only a relatively shallow overdeepening (lower Aare valley) occurs north of the outcrop of the Jurassic limestones (Fig. 3). This is despite the fact that, during the Most Extensive Glaciation (MEG), the ice limit was considerably north of the outcrops of the Jurassic limestones.
- Some overdeepenings are located slightly beyond the LGM ice limit (Fig. 3). This, and the frequent remnants of older glaciations in valleys located within the LGM limits, points towards predominantly pre-LGM formation of the larger overdeepings.

- Some of the valleys in the external Foreland cannot be followed towards the Alps due to bedrock swells within the Molasse substratum. There is a zone with relatively few deep glacial valleys in the internal Foreland (Fig. 2).

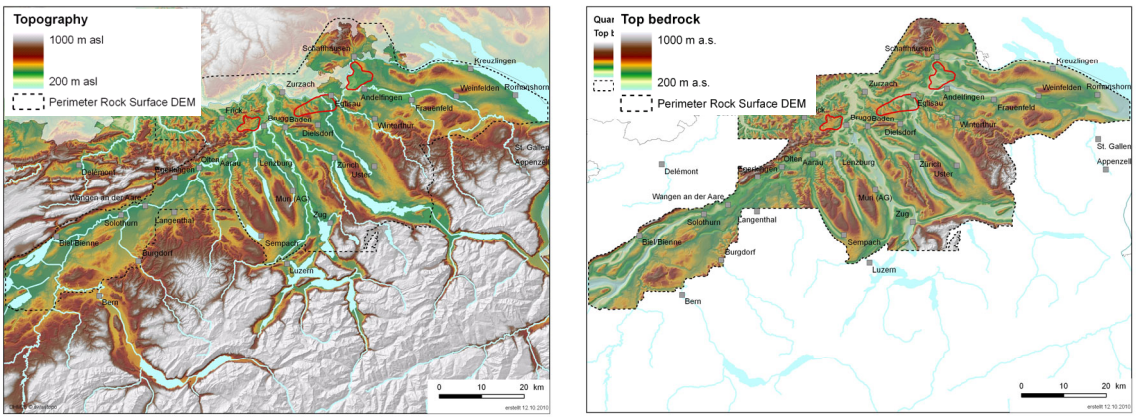

Fig. 2: Digital terrain model (DHM25, swisstopo) and digital top bedrock model of north-eastern Switzerland. Red polygons mark the proposed siting regions under for a high-level waste repository. See Fig. 1 of this report for thickness distribution of Quaternary sediments.

The following questions are of particular relevance for Nagra:

- Can the effect of future glaciations be bounded for specific situations (location, lithology, depth)?
- Where will future overdeepenings occur? What favours the formation of new valleys? What transient effects have to be accounted for (e.g. effect of changing internal Alpine topography during successive glacial cycles on flow and erosion in the Foreland)?
- What is the effect of lithology? To what degree will future overdeepenings cut into Jurassic limestones?

To better understand landscape evolution and better define future erosion scenarios, Nagra is planning to assess and follow state-of-the-art research and to support or initiate specific work concerning (1) process understanding of deep glacial erosion (e.g. through numerical modelling) and (2) the local glaciation history of northern Switzerland (improved characterisation of valley morphology and infill, additional dating of terraces and valley fills).

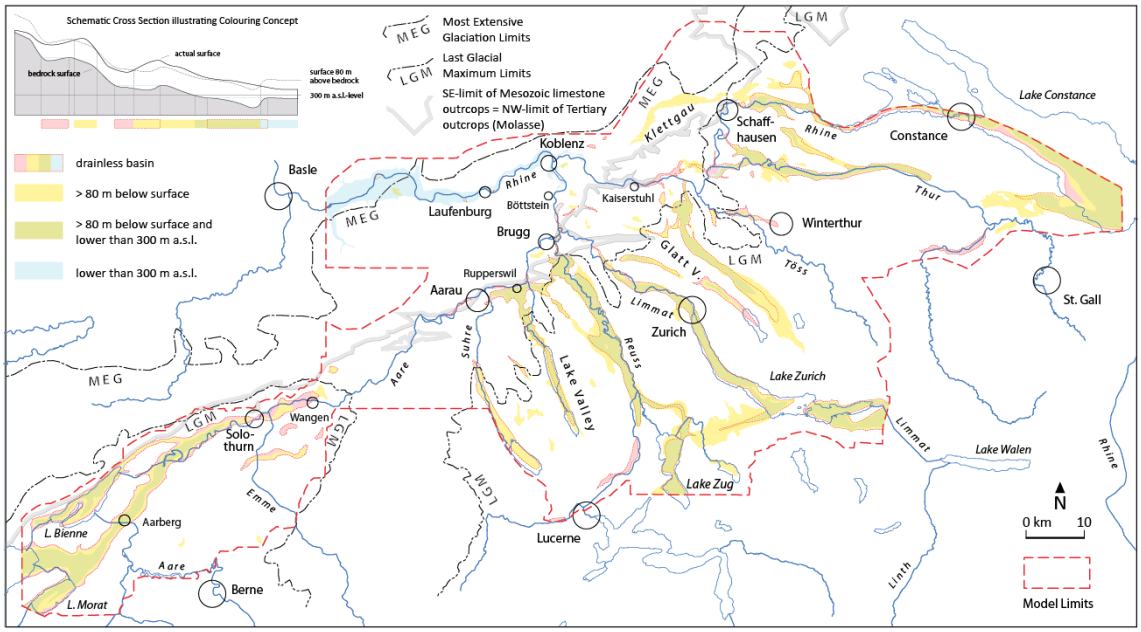

Fig. 3: Analyses of bedrock surface DEM with respect to deep glacial erosion (Jordan 2010). Areas with more than 80 m of unconsolidated rock cover are marked in yellow and green and areas below 300 m a.s.l. are coloured in blue. Undrained (overdeepened) basins are framed by red dots. See cross-section on the top left for an additional explanation of the colouring concept.

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Controls on the efficiency of basal sediment evacuation by subglacial meltwater and the implications for rates and patterns of glacial erosion

Darrel A. Swift

The efficiency of basal sediment evacuation by the subglacial drainage system is likely to strongly limit the efficacy of glacial erosional processes. Sediment transport data from Haut Glacier d'Arolla, Switzerland, demonstrate that a key control on this efficiency is the extent to which a hydraulically efficient channelized subglacial drainage network evolves during the melt season (Swift et al. 2002; 2005a, b; Figs. 1 and 2). For the period prior to evolution of the channel system, suspended sediment concentration (*SSC*) and discharge (*Q*) in the proglacial stream demonstrate that $SSC \propto Q^{1.3}$ (subperiods 1–4). This relationship is consistent with sediment being entrained largely within a hydraulically inefficient (e.g. linked-cavity) drainage system, where changes in discharge are accommodated by changes in both flow velocity and cross-sectional area (e.g. by the enlargement of subglacial cavities). However, from the onset of evolution of the channelized system (around the beginning of subperiod 5), the data demonstrate that $SSC \propto Q^{2.2}$, and that this relationship was sustained even after the system had ceased to extend upglacier (i.e. subperiod 8). This relationship is consistent with sediment being entrained within a partially ice-walled channel system, where rapid diurnal changes in discharge cannot be accommodated by changes in cross-sectional area (e.g. by melting of the channel walls) and hence are accomplished largely by changes in flow velocity (Alley et al. 1997).

The main implication of this change in drainage system morphology was that total daily sediment evacuation during the melt season increased by ~ 125 x, whereas total daily discharge increased by only ~ 7.5 x. Nevertheless, changes in the intercept of the relationship between *SSC* and *Q* between subperiods of the melt season indicate a short-lived decrease in sediment availability during inception of the channelized system, followed by a consistent longer-term increase in sediment availability that was extremely important in maintaining the very high rates of sediment evacuation (in excess of 500 t d^{-1}) towards the end of the melt season. This is also indicated by the pattern in the residuals from the relationship between total daily suspended sediment load (*SSL*) and total daily discharge (i.e. *Residual SSL*; Fig. 1).

The decrease in availability is inferred to be the result of the rapid exhaustion of sediment from the floor of incipient channels, whilst the subsequent increase in sediment availability appears to have been related to the strength of diurnal pressure fluctuations within the channel system (indicated by an absence of pattern in the residuals from the relationship between *SSL* and diurnal discharge amplitude). Hubbard et al. (1995) have demonstrated that such pressure fluctuations result in a diurnally-reversing hydraulic gradient that exists between channels and surrounding areas of hydraulically inefficient drainage that is potentially able to entrain sediment from wide areas of the glacier bed. It therefore seems likely that increasingly strong diurnal pressure fluctuations as a result of the increasing amplitude of the diurnal cycle increased the strength of the diurnally-reversing hydraulic gradient and the entrainment of sediment from increasingly wide areas of the bed. At the same time, a combination of efficient flushing of sediment from the ice-bed interface and enhanced basal sliding during periods of diurnally-high subglacial water pressure may have enhanced the supply of sediment derived from the erosion of subglacial bedrock. These data and observations confirm that the morphology of the subglacial drainage system is a key control on the efficiency of sediment evacuation from the ice-bed interface and indicate that the presence of hydraulically efficient channelized subglacial drainage could influence significantly the overall rate and spatial pattern of erosion and subglacial landscape evolution.

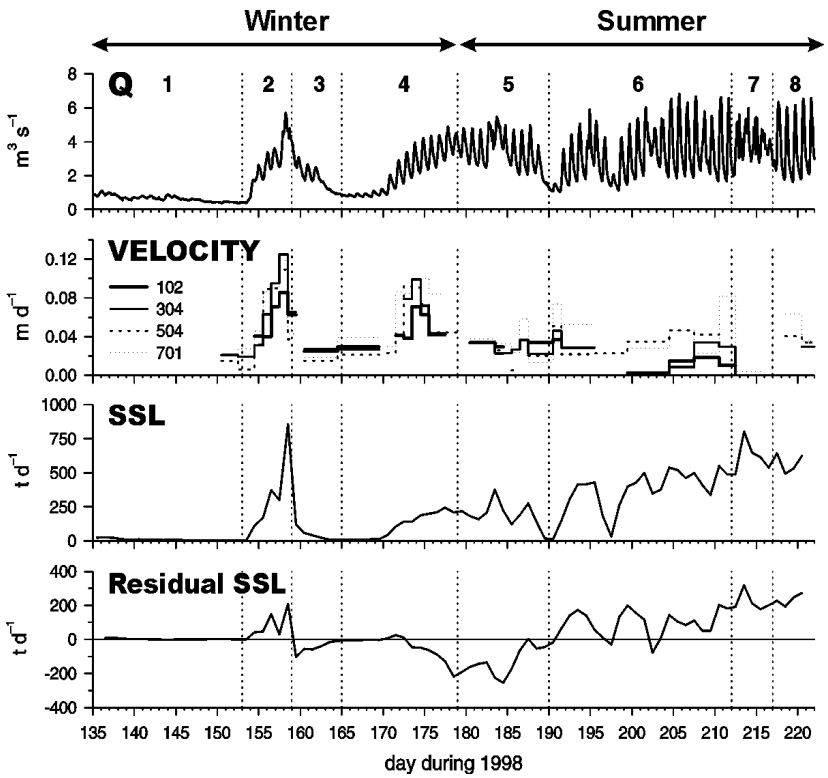

Fig. 1: Hourly mean discharge (*Q*), daily ice-surface velocities and daily suspended sediment load (*SSL*) from Haut Glacier d'Arolla during the 1998 melt season (*Residual SSL* is explained in the text) (Swift 2006).

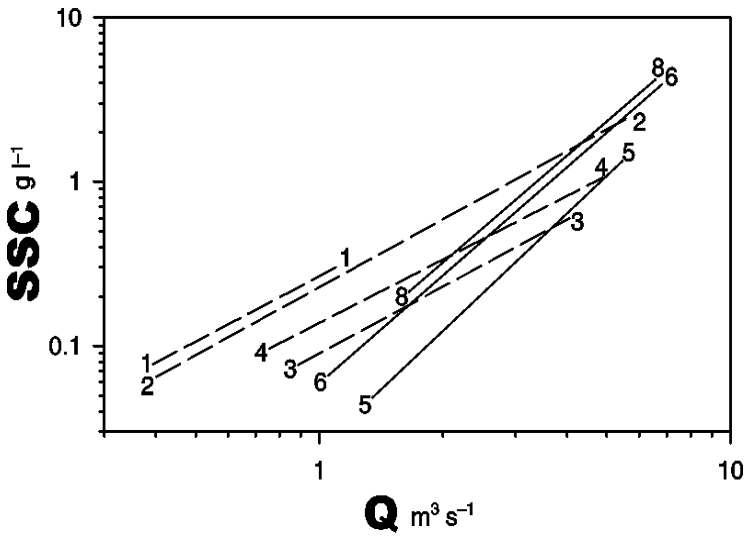

Fig. 2: Relationships between hourly mean discharge (*Q*) and hourly mean suspended sediment concentration (*SSC*) for individual subperiods of the melt season (shown in Fig. 1). Subperiod 7 was dominated by heavy rainfall and is not shown in the graph (Swift 2006).

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Water flow speed in the channelised drainage system

Mauro A. Werder, Thomas V. Schuler, Martin Funk

We present and interpret the results of three series of tracer experiments conducted on an Alpine glacier over a diurnal discharge cycle. Two injection series, using a moulin fed by supraglacial meltwater, yielded observations of tracer transit speed with one diurnal maximum and minimum. The third series, using a moulin fed by a supraglacial lake, produced transit speeds with two diurnal maxima and minima. A simple two-component model of the glacier drainage system, comprising a moulin and a channel element (Fig. 1), was used to simulate the measured transit speeds for the three injection series.

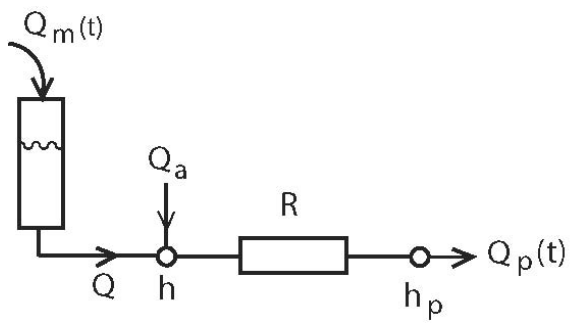

Fig. 1: The lumped element model used for the simulation, consisting of a moulin element (left) and a channel element (right). It is driven by the measured discharge into the moulin and the proglacial discharge and calculates tracer transit speed (Werder et al. 2010).

The model successfully reproduces all the observations and shows that the same underlying processes can produce the qualitatively different behaviour depending on the different moulin input discharge regimes (Fig. 2). The application of the model allows to infer subglacial water flow speeds in the channelised drainage system, which, in terms, should allow to asses the sediment transport capacity of channelised drainage systems.

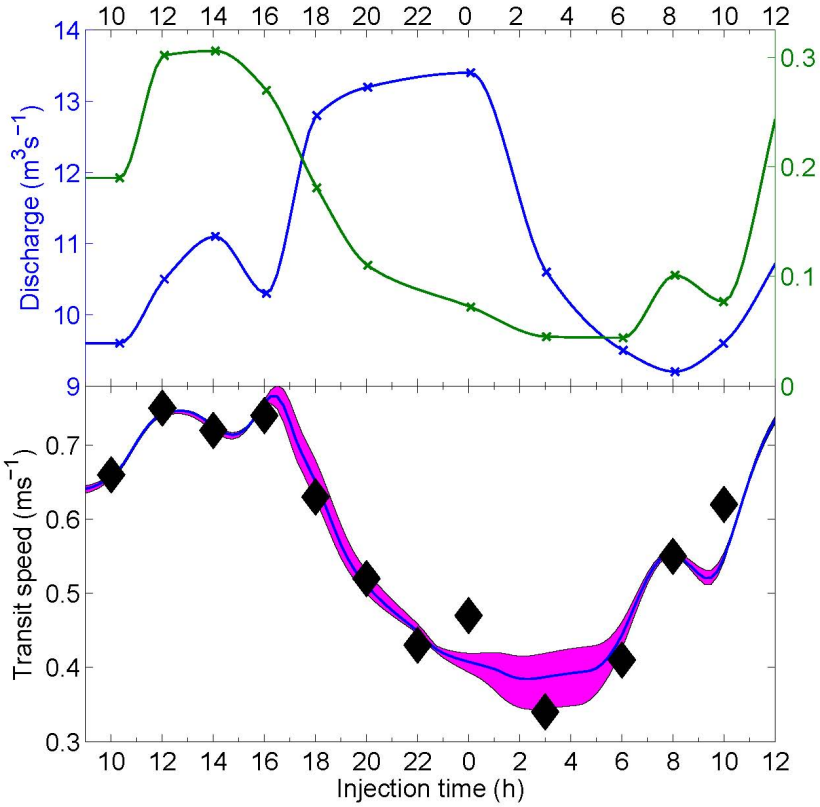

Fig. 2: Upper panel: Measurements of proglacial (blue) and moulin (green) discharge. Lower panel: Tracer transit speed measurements (diamonds) over a diurnal discharge cycle and the results from fitting the model (blue line) with error estimate (magenta) (Werder et al. 2010).

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