Geomorphology 223 (2014) 96–116

Contents lists available at ScienceDirect

Geomorphology

journal homepage: www.elsevier.com/locate/geomorph

# Controls of initial topography on temporal and spatial patterns of glacial erosion

Vivi K. Pedersen [a,*], Ritske S. Huismans [a], Frédéric Herman [b], David L. Egholm [c]

[a] Department of Earth Science, University of Bergen, Allégaten 41, 5007 Bergen, Norway
[b] Institute of Earth Sciences, Université de Lausanne, Bâtiment Géopolis 3871, 1015 Lausanne, Switzerland
[c] Department of Geoscience, Aarhus University, Høegh-Guldbergs Gade 2, 8000 Aarhus, Denmark

ARTICLE INFO

*Article history:*
Received 25 January 2014
Received in revised form 18 June 2014
Accepted 19 June 2014
Available online 10 July 2014

*Keywords:*
Glaciation
Erosion
Landscape evolution
Numerical modelling
Southern Taiwan
Southeast Australia

ABSTRACT

Here we investigate the influence of initial pre-glacial topography on spatial and temporal patterns of glacial erosion using numerical surface process modelling, including a higher order ice sheet model. First, we consider glacier dynamics when simulating glaciation in two real landscapes, representing plateau-type topography (southeast Australia) and characteristic steady-state fluvial topography (southern Taiwan). We find that the different initial landscape configurations result in distinctly different ice configurations and patterns of basal sliding. The sliding patterns are controlled by ice configuration and the resulting basal shear stresses and by the thermal properties at the base of the ice. We then investigate how these characteristic patterns of basal sliding control glacial erosion and long-term landscape evolution using synthetic representations of the two landscapes. The two landscape configurations result in markedly different spatial and temporal patterns of glacial erosion. However, the resulting landscapes may have similar morphology, irrespective of initial landscapes and glacial erosion patterns being significantly different. The numerical experiments also suggest that, in addition to basal temperature, basal shear stress is important in restricting long-term glacial erosion, which is relevant for the preservation of landforms during glaciations. Specifically, pre-glacial landforms may be eroded although they are initially protected by cold-based ice, when the ice configuration promotes significant basal shear stress (glacial erosion) at the edge of a plateau-like landscape. In contrast, pre-glacial landforms may be preserved irrespective of the ice being warm-based, when low gradients in the ice surface act to limit basal shear stress.



## 1. Introduction

Glacial erosion has played a significant role in the recent shaping of high-elevation and mid- to high-latitude mountain regions (e.g., Penck, 1905) and increasing erosion rates worldwide (Zhang et al., 2001; Molnar, 2004; Herman et al., 2013). However, the spatial and temporal variations in glacial erosion responsible for this shaping are generally difficult to unravel on longer time scales (>$10^5$ years), especially as methods constraining glacial erosion rates have proven very dependent on the time scales on which they work (e.g., Gardner et al., 1987; Koppes and Montgomery, 2009). One example of this is found in the highly glaciated St. Elias Mountains, Alaska, where erosion rates have been estimated from sediment yields to ~5–60 mm/y on a millennial time scale (Hallet et al., 1996) and ~5 mm/y since 10,000 years from offshore sediment volumes (Sheaf et al., 2003). On longer time scales (>$10^6$ years) erosion rates are estimated to ~1–5 mm/y based on apatite (U–Th)/He dating (Berger and Spotila, 2008). In addition to intermethod variability, this dependency of erosion rate on the time scale over which it is estimated most likely reflects the response of erosion processes to climate. The climate varies on a range of time scales, including intraglacial, glacial–interglacial, and longer time periods, as a result of orbital forcing, tectonics, and internal dynamics of the climate system (e.g., Zachos et al., 2001; Lisiecki and Raymo, 2005).

Feedbacks between glacial erosion, topography, and the mass balance of glaciers may result in spatial and temporal variations in glacial erosion, irrespective of the climate being constant (e.g., Oerlemans, 1984; Braun et al., 1999; Tomkin, 2003; Kessler et al., 2008; Kaplan et al., 2009; MacGregor et al., 2009; Anderson et al., 2012a; Pedersen and Egholm, 2013; Sternai et al., 2013). On longer time scales, these variations have been proposed mainly based on results from numerical modelling experiments, as long-term erosion estimates from in situ, low-temperature thermochronology often lack temporal resolution (e.g., Shuster et al., 2005; Ehlers et al., 2006; Herman et al., 2013). These methods have provided important constraints on increasing erosion rates at the time of initial glaciations but cannot be used to unravel detailed temporal and spatial patterns in glacial erosion, for example resulting from glacial–interglacial cycles. Recent advances in the methodology of low-temperature thermochronology have improved the temporal resolution of the resulting exhumation patterns, and intra-Quaternary trends in exhumation may be extracted if significant erosion (cooling) has occurred over the evaluated time period (e.g., Shuster

* Corresponding author. Tel.: +47 55 58 34 24.
*E-mail address:* Vivi.Pedersen@geo.uib.no (V.K. Pedersen).

http://dx.doi.org/10.1016/j.geomorph.2014.06.028
0169-555X/

et al., 2011; Valla et al., 2011, 2012). However, tectonically inactive regions such as passive margins pose a challenge when estimating long-term patterns in glacial erosion, as thermochronological ages are usually too old to resolve the most recent Quaternary evolution (e.g., Gallagher et al., 1998; Hendriks et al., 2007; Herman et al., 2013).

Numerical surface process models, including a glacial component, allow for examination of spatial and temporal patterns in glacial erosion, based on a prescribed set of assumptions related to climate forcing, ice dynamics, erosion rules, and topography (e.g., Egholm et al., 2009). While the validity/quality of these assumptions may vary, the approach does enable investigations of high-resolution spatial and temporal patterns of glacial erosion (e.g., Pedersen and Egholm, 2013; Sternai et al., 2013), although providing little constraints on absolute values of erosion on its own. Much work is being done in order to improve the assumptions related to ice dynamics, glacial hydrology, and erosion rules (MacGregor et al., 2009; Egholm et al., 2011, 2012a,b; Herman et al., 2011; Iverson, 2012; Creyts et al., 2013), and in order to understand the Quaternary climate forcing (e.g., Clark et al., 1999; Lisiecki and Raymo, 2005; Huybers, 2006) and how this is reflected in the mass balance of ice bodies (e.g., Braithwaite, 1995; Benn and Lehmkuhl, 2000; Oerlemans, 2002). However, the influence of topography has been overlooked, most likely owing to the generally inadequate constraint on pre-glacial topography.

The choice of initial topography for such numerical forward-model experiments has often been limited either to an existing present-day landscape or a steady-state fluvial configuration (e.g., Herman and Braun, 2008; Egholm et al., 2009; Yanites and Ehlers, 2012; Pedersen and Egholm, 2013; Sternai et al., 2013). However, ice flow, glacier sliding, and glacial erosion are expected to depend strongly on topography. Consequently, limiting the initial topography in forward fluvial–glacial models to the present-day landscape configuration does not allow for exploring the effect of a changing topography over longer time scales (differential glacial erosion). Using an initial fluvial steady state is similarly restrictive, as steady-state morphology appears very rare, at least following periods of significant climate variability, such as the Quaternary (Whipple, 2001; Zhang et al., 2001; Molnar, 2004; Herman et al., 2013).

Here we examine the effects of characteristic end-member pre-glacial topographies on patterns of glacial erosion using numerical forward modelling experiments. We investigate (i) the control of characteristic steady-state fluvial topography and plateau-type topography on glacial sliding patterns, (ii) long-term evolution of glacially dominated landscapes over multiple glacial–interglacial cycles at high spatial resolution in conceptual representations of these landscapes, and (iii) the sensitivity of the resulting patterns in glacial erosion to changes in temperature, precipitation, and model domain size. Finally, we discuss the implications of the numerical experiments for the preservation of pre-glacial landscapes under ice.

## 2. Methods

The numerical experiments presented in this paper have been conducted using a surface process model that includes fluvial incision, mass wasting processes, periglacial frost cracking, and glacial processes (Egholm et al., 2011, 2012a,b). The model is based on a finite-volume approach on an irregular grid, and the governing equations are solved using explicit integration with Euler's method (Egholm and Nielsen, 2010). Details on the model, especially concerning the importance of the higher order ice dynamics involved, have been described in depth elsewhere (Egholm and Nielsen, 2010; Egholm et al., 2011, 2012a,b). The model domain boundaries are kept at a fixed base level throughout the model runs, and any water or sediment that flows to the boundary nodes is removed at the end of each time step. All models are initiated from ice-free conditions, and a low-relief region is added around the model domains in order to prevent ice from flowing to the edge of the models. Information on the various processes included in the model is provided below. All relevant parameters can be found in Tables 1–2.

**Table 1**
Parameters related to fluvial incision and mass wasting[a]. Erosion is only applied in order to generate initial synthetic landscapes.

| Parameter | | Value unit |
|---|---|---|
| $K_f$ | Stream carrying capacity constant | $0.6 \cdot 10^{-4}$ m$^{1.5}$ |
| $m_t$ | Water discharge exponent for carrying capacity | 1.5 |
| $n_t$ | Slope exponent for carrying capacity | 1 |
| $K_{ef}$ | Fluvial erosion constant | 0.6 m$^{-0.5}$ |
| $m_p$ | Water flux exponent for stream power | 0.5 |
| $n_p$ | Slope exponent for stream power | 1 |
| $K_{hs}$ | Sediment transport constant on hillslopes | 5 m$^2$ y$^{-1}$ |
| $s_c$ | Critical slope for hillslope erosion | 1 |
| $K_{eh}$ | Hillslope erosion constant | 0.2 m$^2$ y$^{-1}$ |

[a] For details on the various parameters see Egholm et al. (2012a).

### 2.1. Fluvial processes and mass wasting

Fluvial sediment transport is governed by a stream power capacity model, in which sediment is picked up or deposited depending on the local relation between suspended sediment load and carrying capacity (Whipple and Tucker, 2002). Fluvial incision follows a stream power law, modified to account for the dual effect of sediment being either abrading tools or protective to the bed (Sklar and Dietrich, 1997; Whipple and Tucker, 2002; Egholm et al., 2013).

The transport of material by mass wasting is handled using a nonlinear diffusion model (e.g., Andrews and Bucknam, 1987; Roering et al., 1999; Montgomery and Brandon, 2002). The nonlinearity is used to represent two modes of mass wasting: (i) below a threshold slope, the rate of change in topography is assumed proportional to the curvature of the topography, representing mass wasting processes on long length and time scales; and (ii) above the threshold slope, the diffusion law represents highly effective mass wasting processes working on short time scales to maintain a critical slope.

Fluvial processes and mass wasting are employed in order to generate the initial landscapes used for the glacial modelling. However, our main focus is on glacial erosion in the subsequent numerical experiments that use these landscapes as initial conditions. We include therefore only mass wasting occurring above a critical slope in addition to

**Table 2**
Parameters related to ice dynamics, glacial hydrology, glacial erosion, climate, and mass balance[b], constant for all experiments.

| Parameter | | Value unit |
|---|---|---|
| $\rho_{ice}$ | Ice density | 910 kg m$^{-3}$ |
| $L_{ice}$ | Latent heat of fusion for ice | 334 kJ kg$^{-1}$ |
| $k_{ice}$ | Thermal conductivity of ice | 2.4 W m$^{-1}$ °C$^{-1}$ |
| $c_{ice}$ | Specific heat capacity of ice | 2 kJ kg$^{-1}$ °C$^{-1}$ |
| $dT_h$ | Atmospheric lapse rate | 6.5 °C km$^{-1}$ |
| $T_a$ | Yearly temperature amplitude | 10 °C |
| $q_b$ | Crustal heat flow | 0.045 W m$^{-2}$ |
| $g$ | Acceleration of gravity | 9.81 m s$^{-2}$ |
| $A$ | Ice flow parameter | $10^{-16}$ Pa$^{-3}$y$^{-1}$ |
| $n$ | Ice flow stress exponent | 3 |
| $A_s$ | Ice sliding constant | $3.8 \cdot 10^{-9}$ m Pa$^{-2}$ y$^{-1}$ |
| $m$ | Ice sliding stress exponent | 3 |
| $T_0$ | Max accumulation temperature | 0 °C |
| $s_a$ | Critical slope for snow avalanching | 0.75 |
| $k_{sm}$ | Ablation slope | 0.7 m yr$^{-1}$ °C$^{-1}$ |
| $K_{s:g}$ | Water exchange from ice surface to bed | 0.01 |
| $\phi$ | Average englacial/subglacial porosity | 0.01 |
| $K_{wg}$ | Average en-/subglacial hydrological conductivity | $10^{-4}$ m s$^{-1}$ |
| $K_{wg}^{max}$ | Max. en-/subglacial hydrological conductivity | $10^{-2}$ m s$^{-1}$ |
| $D_b$ | Thickness of debris-rich basal ice | 10 m |
| $D_s$ | Depth of subglacial deformation | 1 m |
| $k_{sg}$ | Apparent conductivity of the array of debris to ice | $10^{-7}$ m$^2$ Pa$^{-1}$ y$^{-1}$ |
| $K_{ss}$ | Subglacial sediment transport constant by streams | $10^6$ s$^2$ m$^{-4}$ |
| $K_a$ | Subglacial abrasion erosion constant | $2 \cdot 10^{-5}$ y m$^{-1}$ |
| $K_q$ | Subglacial quarrying erosion constant | $2 \cdot 10^{-4}$ |

[b] For details on the various parameters see Egholm et al. (2012a).