DISS. ETH NO. 20560

Constraining glacial reshaping of the European Alps: insights from morphometric analyses, paleo-topographic reconstructions and numerical modeling of glacial erosion

A dissertation submitted to

ETH Zurich

for the degree of

Doctor of Sciences

presented by

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2012
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Abstract

A mountain belt like the European Alps is a near ideal setting for studying the tectonic processes that uplift and deform the Earth’s surface and climate-modulated erosion mechanisms that lower topography. Addressing the complex interactions between tectonics and climate in shaping a mountain belt like the Alps requires the use of a variety of methodologies. This thesis focuses on the effects of glaciation in the recent Alpine history by means of morphometric analyses, paleo-topographic reconstructions and numerical modeling. The present-day hypsometry of the Alps reveals that glacial morphologies are mostly preserved at or above the long term glaciers’ equilibrium line altitude. The present-day topography, however, is more prone to bear information about the most recent stages of glacial erosion, as erosion itself erases evidence of past processes and landforms. In order to achieve a more complete understanding of the effects of glaciation on the topographic evolution of the Alps, significant effort was dedicated to the reconstruction of the pre-glacial Alpine topography. Spatial variations of pre-glacial channel steepness suggest a significant effect of rock erodibility in setting the pre-glacial river network elevations and a potential change in regional isostatic support of the western Alps, for example, by detachment of the European slab. The comparison between the modeled pre-glacial topography of the Alps and the present-day landscape highlights the occurrence of significant glacial erosion in the lower and peripheral parts of Alpine catchments, below the long term glaciers’ equilibrium line altitudes. In addition, glacial erosion seems to have decreased the mean elevation and increased valley scale topographic relief across the entire range, by up to a factor of two. Although these results appears to be at odds with the outcomes of the hypsometric analysis of the present-day Alpine landscape, numerical modeling of glacial erosion on the pre-glacial and modern topography provide an explanation for this apparent discrepancy. Glacial erosion on the pre-glacial and present-day topography, in fact, is different and the overall effect suggests an initial increase of topographic relief followed by enhanced erosion at higher elevations during late phases of glaciation. These stages of glacial erosion are linked
by a headward propagation of glacial erosion through time. Glacial erosion also influences the sediment yield. If no additional cooling trends are superimposed to climate oscillations, a tectonic component of rock uplift of $0.5 \text{ mm yr}^{-1}$ is sufficient to increase the production of sediments, in agreement with measurements of sediment yield from the Alps. In addition, a shift of the period of climate oscillations from 40 kyr to 100 kyr can enhance the sediment yield by extending the time spans of glacial erosion and promoting new transient conditions. This mechanism or a change in regional isostatic support of the Western Alps are likely factors to explain, for example, the discrepancy between the age of the earliest glacial deposits in the northern foreland and in the Po plain.

**Sommario**

Le Alpi Europee sono il contesto ideale per lo studio dei processi tettonici e erosionali alla base della formazione di catene montuose di tipo collisionale. Per comprendere le complesse interazioni tra processi tettonici, che sollevano la topografia e ispessiscono la crosta terrestre, e climatici, che definiscono i meccanismi di erosione e ridistribuzione delle masse, è necessario, tuttavia, utilizzare svariate tecniche. Lo scopo di questa tesi è quello di definire gli effetti che le glaciazioni del Quaternario hanno avuto sull’evoluzione topografica delle Alpi tramite analisi di tipo morfometrico, ricostruzioni paleo-topografiche e modellizzazione numerica di processi di erosione glaciale. L’ipsometria attuale delle Alpi rivela che le morfologie glaciali sono prevalentemente preservate ad una quota corrispondente o superiore a quelle che la Linea di Equilibrio dei ghiacciai ha assunto durante il Quaternario. È tuttavia importante considerare che sulla topografia attuale sono preservate più facilmente le tracce di erosione recente piuttosto che antica, in quanto l’erosione stessa cancella le prove di topografie passate. Una parte importante di questa tesi è perciò dedicata alla ricostruzione della topografia Alpina antecedente alla glaciazione. La morfologia Alpina pre-glaciazione suggerisce che l’erodibilità della roccia è un fattore primario nel definire le quote topografiche. Inoltre, pendenze più elevate del reticolo fluviale pre-glaciazione nelle Alpi Orientali suggeriscono un cambiamento di supporto isostatico in quest’area dovuto, per esempio, alla rottura dello slab Europeo. Il confronto tra la topografia attuale e
quella antecedente alla glaciazione mostra come l’erosione glaciale sia più importante nelle parti più basse e periferiche della catena, ben al di sotto della Linea di Equilibrio dei ghiacciai. L’erosione glaciale, in particolare, sembra aver diminuito l’elevazione media delle Alpi ma aumentato il rilievo topografico alla scala di singole valli, fino a raddoppiarlo. Benché queste conclusioni sembrino essere in contrasto con i risultati dell’analisi ipsometrica, la modellizzazione numerica dei processi di erosione glaciale avvenuti sulla topografia antecedente alla glaciazione e quella attuale fornisce una valida spiegazione a questa apparente contraddizione. I ghiacciai, infatti, erodono in modo significativamente diverso la topografia pre-glaciazione e quella moderna. L’effetto complessivo nell’arco di tutta la glaciazione è quello di un iniziale aumento del rilievo topografico, seguito da erosione a quote più elevate durante gli stadi finali della glaciazione. L’erosione glaciale, inoltre, influenza la produzione di sedimenti. Se le oscillazioni climatiche sono costanti nel tempo, una componente di sollevamento tettonica di 0.5 mm yr\(^{-1}\) è sufficiente ad aumentare la produzione di sedimenti, come effettivamente viene osservato nel record Alpino. Anche un cambiamento del periodo delle oscillazioni climatiche da 40 kyr a 100 kyr può aumentare la quantità di sedimenti prodotti dall’erosione glaciale, in quanto viene esteso l’intervallo di tempo durante il quale si sviluppa l’azione erosiva dei ghiacciai e sono favorite nuove condizioni transitorie alle quali il sistema topografico si deve riadattare.Questo processo, o un cambiamento del supporto isostatico delle Alpi Orientali, sono entrambi fattori plausibili per spiegare le differenti età dei primi depositi glaciali nell’avampaese Alpino e nella pianura Padana.
Chapter 1

Introduction and methodology

1.1 Motivation and aims

The Earth is a dynamic planet in which tectonic and climatic forces continuously interact to create and destroy oceans, basins and mountain ranges. The Earth’s topography is the interface where this interaction takes place and, in some way, it bears information about the tectonic processes that deform and uplift the surface and the climatic processes that erode, sculpt and lower the landscape. When continents collide, the interactions between tectonics and climate becomes even more pronounced, although still enigmatic (Figure 1.1).

Ever since Leonardo da Vinci, who was probably the first to make pioneering considerations about “sea floors raised up to mountain top” (da Vinci, 1519) and the erosive power of rivers, scientists such as Lyell, who first stated that the Earth was not created as we see it (Lyell, 1830), Davis, who proposed the first landscape evolution conceptual model (Davis, 1899), and Barrell, who was among the first to introduce the concept of plate tectonics (Barrell, 1914), have tried to disentangle the maze of causes and effects that govern landscape evolution. Understanding the feedback mechanisms, however, is difficult as it involves several complex processes. For example, growth of mountain ranges affects atmospheric circulation, precipitation regimes and erosion patterns (e.g., Ruddiman and Kutzbach, 1989; Raymo and Ruddiman, 1992), but the redistribution of materials through erosion, transport and deposition modifies the stress regimes within the lithosphere and thus
the evolution of the mountain range (e.g., Molnar and England, 1990; Beaumont et al., 1996; Avouac and Burov, 1996; Willett, 1999; Koons et al., 2002).

Figure 1.1: Illustrative representation of mountain building.

Buoyancy and the motion of the plates are the principal drivers that leads to mountain growth. When two or more continental plates with similar buoyancy collide the crust is shortened and thickened and the surface topography is uplifted (e.g., Argand, 1920). Material from the subducted oceanic lithosphere may also be dragged into the stack of continental nappes and included into the orogen (e.g., Dewey and Bird, 1971), while low-buoyancy materials from the continental crust may be subducted and, during later stages of the orogenesis, re-exhumed (e.g., Chemenda et al., 1996). Throughout the evolution of the orogen, other processes such as slab break-off or mantle convection may yield additional components of rock uplift (i.e., uplift of rocks with respect to a fixed base level) and further raise the Earth’s surface (e.g., Davies and von Blanckenburg, 1995; Jolivet et al., 2009). However, such uplifting mechanisms are constantly countered by erosion. Rivers, hillslope processes and glaciers, according to climate conditions and topographic elevation, are the principal agents to sculpt and dismantle the topography.
Under conditions of invariant rock uplift and erosion, the system naturally evolves towards a topographic steady-state (e.g., Adams, 1985; Howard, 1994; Moglen and Bras, 1995; Willett and Brandon, 2002). As initially low-relief landscapes are uplifted, erosion rates steadily increase over time in response to steepening of river profiles and adjacent hillslopes, further enhanced by orographic precipitation. Eventually erosion rates increase sufficiently to counterbalance the rock uplift rate (e.g., Adams, 1985) leading to statistically invariant topography and denudation rates (Whipple, 2001; Willett and Brandon, 2002), unless erosion is so inefficient or rock uplift rate so rapid that the growth of topography reaches limits imposed by crustal strength (Beaumont et al., 1996; Molnar and Lyon-Caen, 1988). Because they are a natural attractor state, steady state landforms constitute a powerful concept for the exploration of relationships between the height and relief of mountain ranges and such factors as rock uplift rate, climate and rock erodibility (e.g., Howard, 1994; Moglen and Bras, 1995; Whipple and Tucker, 1999). Similarly, if steady state or quasi steady state conditions can be demonstrated to hold in particular field areas, morphometric analyses of landforms can be used to test or constrain certain aspects of landscape evolution models (e.g., Sklar and Dietrich, 1998; Slingerland, 1999; Snyder et al., 2000). Denudational and topographic steady state are often assumed in a wide range of studies, from the interpretation of thermochronologic data in terms of exhumation rates (Braun et al., 2006) or the effect of topography on observed cooling rates (Mancktelow and Grasemann, 1997; Stuwe et al., 1994), to morphometric analysis of a landscape.

However, tectonic and climatic forcing is not constant, and a topographic steady state can be very difficult to obtain, as it certainly is to demonstrate (Herman et al., 2010). Alternating fluvial and glacial conditions, for example, produce erosional patterns that are highly variable in time and space and therefore hardly adjustable rock uplift. In addition, the variation of ice load and erosion rates during glacial and interglacial periods modulates isostatic compensation and thus rock uplift rates. Although it is not yet clear whether topographic steady state in fluvio-glacial systems is possible, recent studies suggest that periodic and cyclic glaciation prevent the occurrence of topographic steady state (e.g., Braun et al., 1999; Tomkin and Braun, 2002; Herman and Braun, 2008).

Glaciations have indeed a fundamental role in the evolution of most high-
elevation and mid- to high-latitude mountain ranges on Earth (e.g., Sugden and John, 1976). The late Cenozoic global cooling trend (Zachos et al., 2001) has accompanied, over the last \( \sim 5 \) Myr, an increase in sediment production (Zhang et al., 2001; Hay et al., 2002) and associated components of rock uplift (Molnar and England, 1990; Mitrovica and Peltier, 1991), beside leading to glaciation of the northern Hemisphere at about 2.5 Myr (Raymo, 1994). The assessment of such an increased sediment yield and its effects on global tectonics has been a main focus of recent research. Glacial processes are often singled out to explain increased sediment fluxes (e.g., Piper et al., 1994; Bowerman and Clark, 2011), exhumation histories (e.g., Shuster et al., 2005) or modern patterns of rock uplift (e.g., Peltier, 2004; Spotila et al., 2004). However, this increase in global sediment yield is too early to be attributed to late Pliocene and Quaternary glaciations only and it is not yet clear what impact glaciations have on integrated erosion rates and associated isostatic adjustments (e.g., Molnar and England, 1990; Montgomery, 2002; Brocklehurst and Whipple, 2002; Raymond, 2004). For example, while large scale topographic analyses suggest that glaciations limit mountain growth by increasing erosion around the long-term glaciers’ Equilibrium Line Altitude (ELA) (e.g., Broecker and Denton, 1989a; Brozovic et al., 1997; Brocklehurst and Whipple, 2002; Egholm et al., 2009; Pedersen et al., 2010; Champagnac et al., 2011), measurements of long-term denudation rates and numerical models support increased erosion of lower reaches and consequent increase of topographic relief (e.g., Shuster et al., 2005; Herman et al., 2011b).

The understanding and quantification of glacial erosion, however, are grounded on knowledge of glacial erosional processes. Erosion by glaciers is conventionally attributed to at least three basic mechanisms: abrasion, quarrying (i.e., the growth of cracks in subglacial bedrock and dislodgement of resultant rock fragments) and subglacial fluvial erosion. While subglacial erosion of bedrock by water might be important locally, it is thought to be volumetrically subordinate to abrasion and quarrying (Drewry, 1986). Abrasion by glaciers is relatively well defined (e.g., Hallet, 1979; Iverson, 1991; Lliboutry, 1994) and considered to be a function of the glaciers’ sliding velocity. Most models (e.g., Harbor, 1992; Braun et al., 1999; MacGregor et al., 2000) assume that glacial erosion mostly occurs by abrasion and that sliding velocities, which influence both the flux of abrasive particles (Boulton,
1. Introduction and methodology

1974; Hallet, 1979) and the force with which these particles are pressed against the bedrock (Hallet, 1979; Iverson, 1990), are the dominant factor in determining glacial erosion rates. Other authors, however, consider quarrying to be more important than abrasion (Boulton, 1979; Drewry, 1986; Iverson, 2002), and this claim is supported by measurements (e.g., Loso et al., 2004) and modeling (Hildes et al., 2004). There are numerous discussions of the relative importance of these erosional mechanisms in the literature (e.g., Flint, 1971; Andrews, 1975; Paterson, 1994), but they all suffer from a lack of data.

In the frame of this thesis, I tried to constrain the effects of glaciation by assessing the tectonic, erosional and climatic processes that are responsible for past and present-day topographies. The major questions that this thesis addresses are:

* How much information about past glacial erosion is retained by the present-day topography? Is it possible to quantify it in an objective way?

* Which parts of an alpine landscape are subject to major glacial erosion?

* Do glacial erosion and associated isostatic adjustment increase or decrease topographic relief?

* Is it possible to reconstruct the topography of a mountain range prior to glaciation? What information about patterns and magnitudes of rock uplift and erosion might this provide?

* How do patterns and magnitudes of glacial erosion evolve throughout glaciation?

* How and to what extent glaciations explain increased sediment fluxes, exhumation histories and modern rock uplift patterns?
In an effort to provide answers to these questions, I focused my research on the European Alps. The Alps are a doubly-vergent mountain range resulting from the collision between the European and African plates (e.g., Argand, 1920; Trümpy, 1973). Although the Alpine architecture is very complex, the major aspects of their evolution are relatively well established. Given the historical and geological record of past glaciation, the Alps are often used as a natural laboratory to investigate how glacial processes affect a collisional orogenic belt. Past research has provided an enormous amount of observations on nearly all aspects of Alpine history, in turn making possible the achievement of the objectives set forth.

### 1.2 Tools and methods

The main tools and techniques that I used to achieve these objectives involve Digital Elevation Models (DEMs), morphometric analyses, numerical modeling and inversions of topographic data.

#### 1.2.1 Digital Elevation Models and morphometric analyses

A number of high resolution DEMs from various space missions (e.g., GTOPO and SRTM) are now easily accessible. These models of the topography of the Earth are derived from high precision measurements and can be used as extensive topographic datasets to investigate the present-day morphology. The vast majority of morphometric parameters and techniques presented in this work base their calculations on DEMs. In addition, DEMs can also serve in numerical models, for instance to explore the geomorphic state of a landform with regard to a given process.

A common way to make objective considerations on the geomorphology of a landscape is to produce a morphometric parametrization of the topography. I produced a number of morphometric estimates, from simple calculations of local and regional slope or relief to more complex analyses of the distribution of elevations (i.e., hypsometry), that were used to derive information on the erosional and tectonic history of the European Alps.
1.2.2 Numerical modeling

During the work of this thesis, I took advantage of a large number of numerical models. Most of these, are Landscape Evolution Models (LEMs) solving partial differential equations that encapsulate the physical response of the landscape to different surface processes. I also contributed to developing these LEMs by improving specific aspects or including new elements of complexity as well as creating new algorithms that address the evolution of a landscape.

Numerical models were used to constrain the answers to the main questions of this thesis, mostly by exploring the response of the landscape to a range of different conditions and parameter values. The principal purpose, however, is not to simply reproduce the observations, but rather to understand the role and relative importance of the processes that govern the evolution of a collisional mountain range.

Figure 1.2: Example of a numerical simulation of glaciation on the western Alps.
Fluvial erosion modeling

I made extensive use of the CASCADE numerical model (Braun and Sambridge, 1997). This model includes a river incision law, calculates the evolution of hillslopes through diffusion and accounts for the effects of horizontal advection.

Glacial erosion modeling

The ICE-CASCADE model (Braun et al., 1999; Herman and Braun, 2008) has been a major tool of my research. This code uses the Shallow Ice Approximation (SIA) (Hutter, 1983) to model the evolution of mountain ranges under glacial conditions. In addition to apply the numerical model to constrain specific aspects of glacial erosion, I also contributed to developing the algorithm which now accounts for subglacial hydrology and sediment transport.

Modeling pre-glacial topographies

I dedicated a large part of my research to develop a new algorithm (presented in the fourth chapter) that reconstructs the topography of a glaciated mountain range prior to glaciation. The algorithm resolves spatially variable channel steepness from modern channel head elevations and provide important information on rock uplift and erosion patterns.

1.2.3 Data inversion

While forward models aim to reproduce the observations, the objective of inversion techniques is to use the observations to ameliorate the estimates of modeled parameters. Inversion methods compare the solution of a forward model for an ensemble of parameters to a set of observations. The procedure is repeated modulating the parameters in order to reach an optimum interpretation of the data. For example, the rock uplift and erosion rates that led to the present-day distribution of elevation are complex and generally poorly constrained in most collisional mountain belts. However, by performing a search within possible values and spatial distributions of rock uplift and erosion rates one may compare the resulting
topographies to DEMs and test the suitability of a particular parametrization to represent the landscape.

1.3 Thesis structure

I will provide an overview of the geodynamic and climatic history of the European Alps in the following chapter. Chapters 3, 4 and 5 constitute the main parts of this thesis and include the principal achievements of my research.

In particular, I report in chapter 3 the hypsometric analysis of the present-day topography of the Alps. This study has not only served to develop a new morphometric parameter, specifically designed to quantify in an objective way the glacial imprint on a landscape, but also to recognize that glacial morphologies on the present-day Alpine topography correspond to the long term glaciers’ ELAs.

Chapter 4 deals with the reconstruction of the pre-glacial topography of the Alps and is, perhaps, the most important study of this thesis. An important result of this work is the calculation of the spatial distribution of the pre-glacial channel steepness which gives information about patterns and magnitudes of rock uplift and erosion during the Quaternary, across the entire mountain belt.

Chapter 5 includes the results from numerical modeling of glacial erosion on the reconstructed pre-glacial and present-day topography of the Alps. A major outcome of this study is that patterns and magnitudes of glacial erosion on the pre-glacial and present-day topography are different, which suggests that glacial erosion evolves through time, also in relation with the evolution of the landform. In addition, results indicate that glacial erosion propagates headward, producing an initial increase of topographic relief followed by incision at elevations corresponding to long term ELAs.

The main conclusions of this thesis are summarized in chapter 6.

Appendix A and B include works that I made in collaboration with colleagues as a non-leading author. These studies are part of the research performed during this thesis and they contribute in a substantial way to the main outcomes and conclusions.
1.4 Published material

The most part of this thesis has been published or accepted for publication in international journals. The hypsometric analysis of the present-day topography of the Alps described in chapter 3, for example, has been published in *Journal of Geophysical Research - Earth Surface*, doi:10.1029/2010JF001823. The reconstruction of the pre-glacial topography of the Alps described in chapter 4 has recently been accepted for publication by the journal *Geology*, while the contents of chapter 5 will be submitted for publication to the journal *Earth and Planetary Science Letters*. Similarly, the contents of the Appendix A have been published in *Earth and Planetary Science Letters, vol.310 pp.498-508* and those of the Appendix B will be submitted to the journal *Quaternary Geochronology*. 
Chapter 2

The European Alps

2.1 Introduction

The Alps are a near ideal setting for studying past and present orogenic topography. The long history of research in the Alps has produced a wealth of data unsurpassed anywhere in the world, and the overall orogenic history is reasonably well understood, although very complex. In this chapter, I want to report some of the principal aspects regarding the formation and evolution of the European Alps, that have been the benchmarks of my research.

2.2 Orogenic history and modern architecture

It has been proposed by several authors (Argand, 1916; Heim, 1922; Staub, 1924) that the causal mechanism for the Alpine orogeny was compressional motion between Europe and Africa. As Europe and Africa drifted away from North America along separate paths, the Tethyan seaway slowly closed. In the framework of the theory of plate tectonics, this would appear to be possible, but the detailed analysis of the orogenic history of the Alpine system in terms of the relative motion of Africa with respect to Europe was not immediately obvious because of at least two problems. First, the history of drift in the Atlantic was not known, and hence the temporal and spatial pattern of relative motion between Africa and Europe could not be determined (Dewey et al., 1973). Second, there are at present a
number of microplates between Africa and Europe, each in motion with respect to all adjacent plates (McKenzie, 1970; Vernant et al., 2010). It has required more than a century of research to decipher the relative motion of all microplates and develop a plate tectonic model that may explain the structures and evolution of the Alps. This is best explained in a number of papers (e.g., Argand, 1916; Heim, 1922; Staub, 1924; Dewey et al., 1973; Beaumont et al., 1996). I only report here some major aspects of interest for this study.

The early orogenic phases (i.e., Cretaceous-Eocene, also referred to as the Eoalpine phases) are characterized by subduction of oceanic and sedimentary units (Beaumont et al., 1996), mostly belonging to the European margin. The continental collision started in the Oligocene and led to crustal shortening with complex and mostly north-west-vergent nappes stacking (Argand, 1916). During the Neogene, the Crystalline Massifs (i.e., Aar, Mont Blanc, Aiguilles Rouges, Belledonne, Ecrins and Argentera) were uplifted (Escher and Beaumont, 1997). This event, also marked by the emplacement of the Periadriatic intrusions (e.g., Schmid et al., 1989), was possibly induced by the entrance of the European continental lithosphere in the subduction zone and/or a slab break-off event (von Blanckenburg and Davies, 1995), and gives the orogen the characteristic doubly-vergent (i.e., towards the north-west and south-east) shape. The formation in the Miocene of the Jura mountains, represents the late compressional events (e.g., Burkhard, 1990; Sommaruga, 1999; Willett and Schlunegger, 2010). The Molasse and Po basins collected the sediments and subsided, storing information about the evolution of Alpine convergence and collision since at least the late Neogene (e.g., Schlunegger, 1999; Muttoni et al., 2003, 2007).

The way oceanic and continental domains involved in the orogenesis are arranged reveals significant along-strike changes in the overall architecture of the Alps (e.g., Schmid et al., 2004; Handy et al., 2010). This also appears from geophysical imaging. Existing seismic refraction, near-vertical and wide-angle reflection data has enabled to map the Moho and the top of the lower crust in the Alpine region (Maurer and Anspor, 1992; Schmid et al., 1996; Lippitsch et al., 2003; Kissling et al., 2006). Towards the central Alps, the lower Adriatic (i.e., paleo-Africa) crust is indented within the Moho and the lower European crust and a narrow collision structure exists under the Tauern Window of the eastern
Figure 2.1: Digital elevation model (3 arcsec SRTM) of the European Alps, with simplified tectonic lineaments and units. Red lines represent major tectonic lineaments. Green lines are the External Crystalline Massifs. Violet lines are the Sesia-Lanzo and Dent Blanche zones (Austroalpine, paleo-african domain). Yellow lines represent the major Periadriatic intrusions (i.e., the Adamello and the Bergell plutons). Blue lines show the Tauern and Engadine windows, belonging to the Penninic, paleo-european domain.

Alps. In addition, high resolution teleseismic tomography reveals the geometry of the lower lithosphere. Although southward subduction of the European plate has been inferred throughout most part of the Alpine orogenesis and indeed imaged below the central and western Alps, at present the Adriatic plate underlies the Eastern Alps and is subducted towards the north-east. Tomographic imaging also reveals that part of the subducted European plate below the western Alps is detached (Lippitsch et al., 2003).

Modern convergence rates across the Alps are $\sim 2$ mm yr$^{-1}$ and are primarily set by the counter-clockwise rotation of the Apulian microplate around a Euler pole located in the Po plain (Calais et al., 2002; Nocquet and Calais, 2004; D’agostino et al., 2008). Consequently, the western Alps do not undergo any contraction. However, the observed rock uplift documented by leveling (Kahle et al., 1997; Schlatter et al., 2005) and GPS (Brockmann et al., 2010) show fast rock uplift in the belt, up to 1.5-2 mm yr$^{-1}$. Such a departure between slow or non-existent convergence and fast rock uplift rates is likely to play an important role in defining
erosion rates and sediment production.

2.3 Geodynamic model

The doubly vergent nature of the Alpine belt (Figure 2.2) was noted early in this century by the Swiss geologist Émile Argand. The investigation of the underlying mechanics of continental collision zones was first made by sandbox experiments (Malavieille, 1984). Subsequently, geodynamic models based on numerical techniques (e.g., Willett et al., 1993; Beaumont et al., 1996) have provided a better description of doubly vergent critical wedges (Figure 2.2b).

As I described above, the Alps (and most Alpine-type orogens) are characterized by three distinct tectonic convergent phases: subduction with deformation that has primarily single vergence, a transition from subduction to collision and continental collision with double vergence. Although the Cenozoic history of the European Alps has additional complexities, a mechanical explanation for these three phases provides the necessary crustal-scale framework to describe a double vergent structures. Subduction within the mantle lithosphere, in fact, leads to the development of two major oppositely dipping shear zones in the crust (Willett et al., 1993; Beaumont et al., 1996). On this model, known as the doubly vergent critical wedge model (Koons, 1994), it is possible to define the pro- and retro-shear zones (Willett et al., 1993), developing at approximately 45° in the opposite direction. In the European Alps, these two structures are comparable in scale and position with the Penninic front and the Periadriatic line, respectively. The position of these zones, where strain accumulates, is thought to stay unchanged as the model develops, but rocks are displaced horizontally and vertically through these features towards the center of the wedge. The style of crustal deformation is very similar to that proposed by Argand (1916) almost one century ago.

2.4 Climatic history

Through the study of sedimentary archives, it has become increasingly apparent that, during much of the last 65 million years, Earth’s climate system has experi-
2. The European Alps

Figure 2.2: a) Conceptual model of the structural evolution of the Alps by E. Argand (1916). b) Results from numerical modeling of the mechanical evolution of a subduction-collision system by Beaumont et al. (1996). Note the similarities between the final stage in a) and the "phase 3" stage in b).
enced continuous change, drifting from extremes of expansive warmth with ice-free poles, to extremes of cold with massive continental ice-sheets and polar ice caps (Zachos et al., 2001). Such change is not unexpected, because the primary forces that drive long-term climate, Earth’s orbital geometry and plate tectonics, are also in perpetual motion. Much of the higher frequency change in climate ($10^4$ to $10^5$ years) is generated by periodic and quasi-periodic oscillations in Earth’s orbital parameters of eccentricity, obliquity and precession that affect the distribution and amount of incident solar energy (Milankovitch, 1969). The finer details of the evolution of global climate over the Cenozoic have come to light thanks to investigations of deep-sea sediment cores (e.g., Ruddiman and Kutzbach, 1989; Raymo and Ruddiman, 1992; Zachos et al., 2001; Lisiecki and Raymo, 2007). Since the mid-Pliocene the climate has become both colder and more variable, with large amplitude temperature fluctuations reflecting both eccentricity and precession forcing (Figure 2.3). This cooling trend led to the onset of Northern Hemisphere glaciation at about 2.5 Myr (Raymo, 1994).

At the beginning of the twentieth century, the classical fourfold system for the glaciation of the Alps was presented by Penck and Brückner (1909). Distinct morphostratigraphic elements, for the most part terraces, moraines and erratic boulders, located on the northern forelands of the Alps were the basis for setting up the following system: (1) Günz, as defined by older (or higher) ”Deckenschotter” (cover gravel) deposits, (2) Mindel, as defined by younger (or lower) Deckenschotter deposits, (3) Riss, defined by moraine remnants and outwash related to the high terraces (Hochterrasse) and (4) Würm, delineated by moraines and outwash of the low terraces (Niederterrasse). The Riss and the Würm were separated by the last interglacial. The system was subsequently expanded and repeatedly revised. Buoncristiani and Campy (2004), Schlüchter (2004), van Husen (2004) and Ivy-Ochs et al. (2009) summarize the present state of knowledge. Terminal moraines and associated deposits formed during the Last Glacial Maximum (LGM, late Würm) provide the most important morphological reference point in the forelands of the Alps (Figure 2.4). The final disintegration of the foreland piedmont glaciers marks the start of the “Alpine lateglacial” as defined by Penck and Brückner (1909). During this time, glaciers re-advanced several times to successively smaller extents (“stadials”), building prominent moraine systems in the valley and cirques.
2. The European Alps

Figure 2.3: Global deep-sea oxygen and carbon isotope records based on Zachos et al. (2001). The sedimentary sections from which these data were generated are classified as pelagic (e.g., from depths >1000 m) with lithologies that are predominantly fine-grained, carbonate-rich (>50%) oozes or chalks. Most of the data are derived from analyses of two common and long-lived benthic taxa, Cibicidoides and Nuttallides. The $\delta^{18}$O temperature scale was computed for an ice-free ocean, and thus only applies to the time preceding the onset of large-scale glaciation on Antarctica (∼35 Myr).
2.4 Climatic history

There is evidence of an increase in sediment accumulation in oceanic and continental basins over the last 5 Myr (Hay et al., 1988). Although some have questioned these effects (e.g., Willenbring and von Blanckenburg, 2010), it has been proposed that global cooling has increased erosion rates in most of the world’s orogenic belts (Hay et al., 2002; Molnar, 2004; Zhang et al., 2001). This increase in sediment yield, however, is too early to correspond exclusively to Quaternary glaciation. Estimates of long-term sediment yield in the Alps are consistent with global observations (Hay et al., 1992; Kuhlemann, 2000; Hay et al., 2002), but the earliest glacial deposits are as old as ∼ 2 Myr (Schlüchter and Kelly, 2000; Ivy-Ochs et al., 2009) and glacial deposits in the Po plain suggests an even later onset (or intensification) of glacial erosion at ∼ 0.9 Myr (Muttoni et al., 2003). The entire Plio-Pleistocene is a period of rapid climatic change, at least globally, so this increase in sediment flux could still be climatically driven, but if so, it is not exclusively a glacial erosion signal. In any case, the correlation between the sediment yield curve (Figure 2.5) and global temperature proxies has stimulated much recent interest and research.

Figure 2.4: Digital elevation model (3 arcsec SRTM) of the European Alps with the Last Glacial Maximum (Würm) ice extent as compiled by Ehlers and Gibbard (2004) (white line). Dashed and dotted lines represent the main and secondary water divides, respectively.
2. The European Alps

2.5 Geomorphology

The combined effects of tectonics, rivers and glaciers is recorded in the geomorphology of the Alps. The overall shape of the mountain chain (Figure 2.4) is characterized by a narrow, N-S trending western part and a wider, E-W trending segment towards the east. Although variations of tectonic processes across the Alpine arc (e.g., Calais et al., 2002; Lippitsch et al., 2003; Nocquet and Calais, 2004; D’agostino et al., 2008) are an important factor in setting major morphologic features, there might also be some erosional controls to these large scale differences between western and eastern Alps. For example, a decrease of erosional efficiency in the early Miocene, coinciding approximately with widespread exposure of the crystalline core in the Alpine hinterland, might have initiated a transition from a phase dominated by vertical extrusion to a period dominated by lateral orogen growth (Schlunegger and Simpson, 2002). The Alpine drainage system, whose principal architectural elements were established during the Neogene (e.g., Schlunegger et al., 2001; Schmid et al., 2004), is characterized by four major rivers draining the orogen towards the north (Rhine and Danube systems), west (Rhone system) and south (Po system) (Figure 2.4).

Although the Alps have originated and evolved mostly under fluvial conditions,
2.5 Geomorphology

the exceptionally high topography and relief that characterizes the present-day landscape is presumably due to the glacial overprint (e.g., Penck, 1905; Haeuselmann et al., 2007; Valla et al., 2011). Major Alpine valleys were occupied by large glaciers, which have widened and overdeepened them (e.g., Penck, 1905; van der Beek and Bourbon, 2008; Norton et al., 2010). Glacially overprinted valleys were subsequently filled by late- and post-glacial lakes and sediments (Hinderer, 2001; Preusser et al., 2010). The present-day morphology of the Alps and the foreland basins is dominated by erosional and depositional features related to
late-Pleistocene and Quaternary glaciations (Kelly et al., 2004; van der Beek and Bourbon, 2008). Valley longitudinal profiles often present a succession of steps and flats (e.g., van der Beek and Bourbon, 2008; Valla et al., 2010b), with valley bottoms at around \( \sim 1000 \) m and summits over 4000 m elevation. Towards the peripheral part of the belt, glacially overdeepen troughs are numerous (e.g., Preusser et al., 2010) (Figure 2.6). In the inner parts, where post-glacial sedimentary in-fill or lakes do not cover the bedrock, the main trunk and tributary valleys show characteristic glacial “U”-shapes. At the confluences of small tributary valleys with a main through-going valley, hanging valleys are also common (e.g., Bini and Zuccoli, 2004; Bini et al., 2007). Such morphologic features are consistent with a glacial erosion model that links confluence steps and valley overdeepenings to the relative erosive power of the joining glaciers (MacGregor et al., 2000; Anderson et al., 2006).

Hanging valley terminations are commonly marked by free-falling waterfalls or bedrock gorges (Figure 2.7, e.g., Valla et al., 2010a). The depth of these gorges is such that fluvial erosion rates of \( \sim 8.5-18 \) mm yr\(^{-1} \) would be required to create their current relief solely during the present interglacial period (Montgomery and Korup, 2010). Such rates exceed the long-term average bedrock erosion rates of even the most tectonically active regions. Therefore, Alpine gorges are more likely to be progressively incised below the elevations of glacial trough valleys through multiple glaciations. These gorges are often associated with narrow and steep bedrock channels with dominant step-pool and boulder-cascade bed morphologies. Blocks derived from the gorge sidewalls or surrounding slopes are frequent and suggest important hillslope-channel coupling (Korup and Schlunegger, 2007).
Figure 2.7: Field photos from Valla et al. (2010b) showing glacial morphologies and bedrock gorges from the western Alps. (a) View of a small glaciated catchment with a glacial hanging valley that ends in a waterfall (Cascade de la Pisse, Vénédon valley). (b) Detail of a gorge waterfall incising the sidewall of the trunk glacial valley (Saut de la Pucelle, Gâ gorge, Upper Romanche valley). (c) Confluence between hanging and trunk valley showing incised “U”-shaped glacial hanging valley form (Vallon stream, Vénédon valley). (d) Bedrock gorge channel, filled with meter-scale blocks, exhibiting step-pool morphology (Diable gorge, Vénédon valley).
Chapter 3

Hypsometric analysis to identify spatially variable glacial erosion


3.1 Introduction

Since Molnar and England (1990) posed the “chicken or egg” conundrum, numerous studies have attempted to identify the tectonic response to climatic variability, and conversely whether changes in tectonics can influence climate (e.g., Raymo and Ruddiman, 1992; Brozovic et al., 1997; Whipple et al., 1999; Reiners et al., 2003). While most research has focused on the role of fluvial erosion on the topographic evolution of mountain ranges (e.g., Koons, 1990; Beaumont et al., 1992; Avouac and Burov, 1996; Willett, 1999; Willett et al., 2001), the effects of glacial erosion remain unclear (e.g., Adams, 1980; Whitehouse, 1987; Brozovic et al., 1997; Whipple et al., 1999; Tomkin, 2000; Brocklehurst and Whipple, 2002; Shuster et al., 2005; Herman and Braun, 2006; Herman et al., 2007, 2010). As glaciers have been prevalent throughout much of the Plio-Pleistocene, the understanding of their ability to modulate erosion rates is of key importance (Zhang et al., 2001; Willenbring and von Blanckenburg, 2010).
3.1 Introduction

Qualitative observations suggest that glacial erosion is maximized around and above the mean long term Equilibrium Line Altitude (ELA) (e.g., Brocklehurst and Whipple, 2004; Anderson et al., 2006; Mitchell and Montgomery, 2006), meaning an ice controlled limit on peak elevation, independent of rock uplift rate, termed the “glacial buzz-saw” hypothesis (e.g., Brozovic et al., 1997). These observations are supported by global scale analyses of the distribution of elevations that show a strong correlation between variations in peak elevation and climate-controlled snowline altitude (e.g., Porter and Zhisheng, 1995; Egholm et al., 2009; Pedersen et al., 2010). Elevation distribution maxima towards an altitude window just below the snowline, consistent with the glacial buzz-saw hypothesis, can in general be explained by a combination of glacial erosion mechanisms such as abrasion, quarrying and subglacial fluvial activity (e.g., Boulton, 1979; Hallet, 1979; Iverson, 1991, 1999). However, although the signatures of these glacial erosion mechanisms, such as “U”-shaped valley bottoms, hanging valleys, overdeepened bedrock basins and glacial cirques, show significant departure from the morphology of fluvial systems (e.g., Penck, 1905; Flint, 1957), the objective quantification of such glacial erosion on topography is still difficult.

Hypsometry (i.e., the frequency distribution of elevations) has been employed to resolve spatially variable uplift and erosion rates of a landscape and whether the landform is characteristic of fluvial or glacial processes (e.g., Strahler, 1952; Montgomery et al., 2001; Brocklehurst and Whipple, 2004; Walcott and Summerfield, 2007; Pérez-Peña et al., 2009; Pedersen et al., 2010). However, it is often not obvious how to interpret the results of such an analysis as the response of hypsometry to tectonic and climatic variability is not unique.

In this study, we provide an overview of hypsometric analysis with a particular focus on the use of the hypsometric curves and integrals to quantify the glacial imprint of topography. Through the study of synthetic cross-sectional valley profiles we describe the major changes that glacial erosion induces on the distribution of elevations. We derive from the hypsometric curves of ideal topographies an additional morphometric parameter, termed the hypokyrtome, defined as the minimum normalized elevation at which the gradient of the hypsometric curve is greater than or equal to a reference value. We first apply the hypsometric analysis to the Ben Ohau Range within the Southern Alps, New Zealand, where the erosive
3. Hypsometric analysis to identify spatially variable glacial erosion

history has been established from geomorphic mapping (Kirkbride and Matthews, 1997). We then use a landscape evolution model to track the morphologic evolution of regions undergoing different tectonic and climatic conditions. Finally, we apply the analysis to the European Alps and the northern Italian Apennines, which have experienced extensive and limited glaciations respectively (e.g., Penck, 1905; Jaurand, 1994; Muttoni et al., 2003; Giraudi, 2004; Ivy-Ochs et al., 2009).

3.2 Hypsometric analysis to characterize fluvial and glacial landscapes

In this section, we first analyze the differences that may arise between hypsometric curves and hypsometric integrals computed on fluvial and glacial topographies. Note that, for convenience, we consider idealized landscapes although it is clear that reality is more complex. We then introduce the hypsokyrtome and present some preliminary considerations about the applicability and efficacy of the hypsometric analysis to investigate the degree of glaciation on mountain ranges.

3.2.1 Hypsometric curves and integrals

Hypsometry is often computed using the hypsometric curve, i.e., the cumulative histogram of elevations within a catchment (e.g., Strahler, 1952; Brocklehurst and Whipple, 2004) or region (e.g., Pérez-Peña et al., 2009; Egholm et al., 2009). The hypsometric integral is the area lying below the normalized hypsometric curve. Historically, Strahler (1952) defined the hypsometric integral, $HI$, as follows

$$HI = \int_0^1 h' da'$$

where $h' = \frac{h - H_{\text{min}}}{H_{\text{max}} - H_{\text{min}}}$, $h$ (m) is the elevation, $H_{\text{max}}$ (m) is the maximum elevation and $H_{\text{min}}$ (m) is the minimum elevation, and $a' = \frac{a}{A_{\text{max}}}$, where $a$ (m$^2$) is the area above $h$ and $A$ (m$^2$) is the total area. In the following, we adopt the same approach as Brocklehurst and Whipple (2004) and Egholm et al. (2009), where $a$ is instead the area below $h$. We then calculate $HI$ as:
3.2 Hypsometric analysis to characterize fluvial and glacial landscapes

\[ HI = \frac{1}{n} \sum_{i=1}^{n} h'_i \]  

(3.2)

where \( n \) is the number of bins used to discretize the elevation range.

![Diagram of hypsometric curves](image)

Figure 3.1: Illustrative representation of hypsometric curves referring to cross-sectional profiles of synthetic valleys. a-c) Profiles and hypsometric curves of synthetic topographies characterized by a “V”-shaped valley bottom. d-f) Profiles and hypsometric curves of synthetic topographies characterized by a “U”-shaped valley bottom. \( HI \) for each curve is: \( HI_a \approx 0.5; HI_b \approx 0.4; HI_c \approx 0.7; HI_d \approx 0.2; HI_e \approx 0.3; HI_f \approx 0.5. \) Note that the hypsometric curves a) and f) have the same \( HI \).

Recently, Egholm et al. (2009) analyzed the hypsometry of glaciated mountain ranges on a global scale. Their results clearly illustrate how hypsometry evolves in response to glacial erosion as well as the relationship that exists between the hypsometric maxima, snowline altitudes and glaciers’ ELAs. Similarly, Brocklehurst and Whipple (2004) showed that, since glacial valley floors represent shallow slopes in the topography, they appear in the hypsometric curve as reaches where there is a high proportion of the area within a modest elevation range. Brocklehurst and Whipple (2004) also argued that \( HI \) can provide information about the amount
3. Hypsometric analysis to identify spatially variable glacial erosion

of glacial erosion. It indeed seems reasonable to compare fluvial landscapes to
glacial ones if one assumes that the cross-sectional profile through an idealized
fluvial valley is represented by a “V”-shape and a glacial one by a “U”-shape. Fig-
ures 3.1a and 3.1d show that these end-member cases lead to distinct hypsometric
curves and integrals. However, slightly more complex, transient morphologies may
generate inflection points on the hypsometric curve (e.g., Figures 3.1b, 3.1c, 3.1e,
3.1f), which in turn lead to changes in the value of $HI$. Furthermore, completely
different hypsometric curves can have the same $HI$ (Figures 3.1a and 3.1f). While
our simplified theoretical 2-D cross-sectional profiles on Figure 3.1 do not provide
an exhaustive description of the morphologic changes induced by erosional pro-
cesses, they highlight the inherent difficulties of using $HI$ as a means to quantify
the influence of glacial erosion on landscapes.

3.2.2 The Hypsokyrtome

Following the discussion in the previous section, our goal is to derive from the
hypsometric curve an additional morphometric parameter that scales with glacial
erosion. It seems reasonable to assume that, during the onset of glaciation, most
parts of the landscape evolve from “V”- to “U”-shaped valleys. Numerical simu-
lations of glacial erosion on valley cross-sections (e.g., Harbor et al., 1988; Harbor,
1992) suggest initially higher erosion rates in the lower parts of the valley, towards
the center of the glacier, where the ice thickness and sliding velocity increase.
As the model evolves, constrictional stresses due to the valley flanks reduce the
sliding velocity, and thus erosion rates, at the center of the “V”-shaped chan-
nel and migrate velocity maxima toward the flanks. Such a velocity distribution
reduces the elevation difference between the valley bottom and surrounding ele-
vations, and leads the valley to progressively evolve from a “V”- to a “U”-shape.
As shown in Figure 3.1 and 3.2, this transition leads mostly to changes in the
slope, $S = \Delta h'/\Delta a'$, of the lowest part of the hypsometric curve (i.e., where $h'$
is small). To characterize this effect, we introduce the hypsokyrtome (from the
Greek: ipsos=elevation, kyrto: curvature), $H_{kr}$, which represents the minimum
$h'$ at which $S$ is greater than or equal to a reference value:
3.2 Hypsometric analysis to characterize fluvial and glacial landscapes

\[ H_{kr} = \min (h'(S \geq S_f)) \]  

where \( S_f \) must be chosen as discussed below.

Figure 3.2: Illustrative representation of glacially induced widening of an ideal “V”-shaped cross-sectional valley profile (a), and the consequent evolution of the hypsometric curves and \( H_{kr} \) (b). The elevations are normalized thus glacially induced valley deepening is not represented in a). This figure is built with respect to Figures 3.1a and 3.1d and Figure 3.1a has been used to compute \( S_f \), it thus equals 1. Light gray represents the early stages whereas dark gray is used to illustrate an extensively glaciated topography. Dashed lines in a) and b) mark the modeled ice surface and \( H_{kr} \) respectively. The stars in b) mark the point where \( S = S_f \). Note that this figure can also be interpreted as the illustrative representation of the cross-sectional valley profiles and hypsometric curves of different parts of the same catchment, where light and dark gray is used for lower (i.e., fluvial-dominated morphologies) and higher (i.e., glacial-dominated morphologies) parts of the catchment respectively.

3.2.3 Applying the hypsometric analysis

Hypsometry provides a scale dependent description of landforms that highly depends on the choice of a series of parameters (e.g., Strahler, 1952; Lifton and Chase, 1992; Walcott and Summerfield, 2007). For instance a bin size, \( b_s \), must be selected to build the hypsometric curve. Similarly, \( H_{kr} \) requires selecting \( S_f \), while both \( HI \) and \( H_{kr} \) must be computed within a given region or cell (i.e., window) whose size must be defined. In this section, we discuss how these parameters can be chosen.
3. Hypsometric analysis to identify spatially variable glacial erosion

Choice of the bin size, $b_s$

Potential trends within a DEM can be better recognized when the elevation values are binned (e.g., Strahler, 1952; Lifton and Chase, 1992). Thus, the choice of $b_s$ is of critical importance. Defining relief, $R$, as the difference in elevation between the highest and the lowest altitude, $b_s$ is determined by dividing the relief by the number of desired bins. If $b_s$ is too large, the hypsometric curve will be too rough (i.e., with increased variance) due to the inherent roughness observed in topography and additional noise within digital elevation models. On the other hand, if $b_s$ is too small, then the hypsometric curve will be too smooth with a potentially large bias. Therefore, the choice of $b_s$ should balance the trade-off between variance and bias. Although there is no optimal approach to choose $b_s$, one may use the approach of Scott (1979): $b_s = 3.5\sigma/(n_p^{1/3}R)$, where $\sigma$ is the standard deviation of the elevation values and $n_p$ is the number of pixels analyzed. This approach applied to large scale analyses leads to values between 30 m and 100 m, also similar to bin widths used in previous studies of elevation frequency distribution (e.g., Brocklehurst and Whipple, 2004; Egholm et al., 2009).

Choice of $S_f$

$S_f$ is a reference value that represents the gradient of fluvial hypsometric curves. The departure between $S_f$ and the gradient of the hypsometric curve of a study region provides an estimate of the glacial imprint on the region of interest. However, the complexity and variability of fluvial morphologies across different mountain ranges make it difficult to determine a unique $S_f$. We therefore seek values of $S_f$ that are appropriate for the tectonics and climate of a region of interest. Despite the morphologic evidence of glacial erosion that can be found at the lower altitudes of glaciated orogens (e.g., overdeepenings in peripheral valleys and glacial lakes), previous studies (e.g., Brozovic et al., 1997; Brocklehurst and Whipple, 2004; Anderson et al., 2006; Egholm et al., 2009) suggest that the integrated glacial erosion on mountain ranges is maximized around and above the mean long-term ELA. On the basis of these studies, a mountain range can be split into two zones with different erosive regimes: (1) elevations below the mean long term ELA, which were mainly subject to fluvial erosion and (2) elevations around and above the mean
3.2 Hypsometric analysis to characterize fluvial and glacial landscapes

long term ELA, where there has been significant glacial erosion as well as fluvial erosion. The low elevation part, mostly presenting a fluvial morphology, is likely to have undergone climatic and tectonic conditions that are consistent with the geological context of the rest of the range. Thus, it can be considered as a suitable fluvial region to compute $S_f$. As glacial and fluvial reshaping of the valley bottom mostly affects the slopes of the lowest part of the hypsometric curve (Figure 3.1 and 3.2), one may define $S_f$ as the average gradient of the lower half of a fluvial hypsometric curve. In Table 3.1, we provide some examples of $S_f$ for a selection of mountain ranges, calculated by using the hypsometric curves of the lower sector (i.e., below the LGM ELA) of each mountain range. $S_f$ appears to range between ~1.5 and ~2.

Table 3.1: Summary of $S_f$ for a selection of mountain ranges. $S_f$ and the standard deviation have been computed by averaging the mean gradient of the lower half of the hypsometric curves of the part of the landscape below the LGM ELA for the region of interest (*Benz, 2003; Buoncristiani and Campy, 2004; Giraudi, 2004), **(Porter, 1975), ***(Brocklehurst and Whipple, 2004), ****(Cui et al., 2002)).

<table>
<thead>
<tr>
<th></th>
<th>W/E/S/N</th>
<th>LGM ELA (m)</th>
<th>$S_f$</th>
<th>Stdv(1σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Southern Alps, NZ</td>
<td>165/175/-50/-40</td>
<td>1000**</td>
<td>1.60</td>
<td>0.43</td>
</tr>
<tr>
<td>European Alps</td>
<td>5/15/44/48</td>
<td>1000*</td>
<td>1.63</td>
<td>0.57</td>
</tr>
<tr>
<td>Apennines</td>
<td>10/15/40/40</td>
<td>1700*</td>
<td>1.97</td>
<td>0.48</td>
</tr>
<tr>
<td>Euro.Alps+North.Apenn.</td>
<td>5/15/40/50</td>
<td>1000*</td>
<td>1.82</td>
<td>0.54</td>
</tr>
<tr>
<td>Taiwan</td>
<td>120/125/20/30</td>
<td>3500****</td>
<td>1.58</td>
<td>0.51</td>
</tr>
<tr>
<td>Eastern Sierra Nevada</td>
<td>-120/-115/35/40</td>
<td>3000****</td>
<td>1.83</td>
<td>0.54</td>
</tr>
</tbody>
</table>

Choice of the cell size

We follow the same approach as Pérez-Peña et al. (2009) and Egholm et al. (2009) by using a grid of squared cells for the computation of the hypsometric curves, rather than drainage basins (Brocklehurst and Whipple, 2004). This choice enables us to avoid the influence of the basin outlet geometry, which controls the shape of the lower part of the hypsometric curve of a catchment and, therefore, the choice of $S_f$. This approach requires defining an optimal cell size that must be consistent with the main wavelength of the topography and the aims of the analysis. As
the main objective of our study is the quantification of the glacial signature on the topography of mountain belts, the cell must cover enough area to be able to isolate glacial morphologic features from fluvial ones, but it should also be sufficiently small to resolve the effects of glacial reshaping on the hypsometric curve (as described in section 3.2.2 and Figure 3.2). In geostatistical studies, an empirical variogram, $2\gamma$, is commonly used to indicate correlation within data (e.g., Curran, 1988; Delay and de Marsily, 1994). It is defined as the expected squared variance of the observations at different measurements (Armstrong, 1984) and it is commonly represented as a graph that shows variance against distance of a measure (e.g., Hamlett et al., 1986; Sen, 1989; Wackernagel, 2003). Here, we look at the correlation in the value of the $H_{kr}$ computed for different cell sizes. In particular, we compute the squared variance of the mean $H_{kr}$, measured for different cell sizes over the study region:

$$2\gamma(k) = \frac{N}{N} \sum_{i=1}^{N} (H_{kr}^{(c+k)}(i) - H_{kr}^{(c)}(i))^2$$

(3.4)

where $k = 0, 1, 2, ..., n$ (pixels) is the increment in the length of the cell, $c = 25$ (pixels) is the minimum cell length and $N$ is the number of pixels included in the study region. The optimal cell size is reached when the squared variance is stabilized at a constant value. We obtain cell sizes ranging between $\sim 15 - 30$ km. These cell lengths are most suitable for analysis at the scale of the whole mountain range (also in agreement with previous studies (e.g., Brozovic et al., 1997; Brocklehurst and Whipple, 2004; Pérez-Peña et al., 2009; Egholm et al., 2009)) as the hypsometric distribution becomes significant when the cell size is comparable to the spacing of the main transverse catchments draining an orogen.

### 3.3 Test applications of the hypsometric analysis

#### 3.3.1 Hypsometric analysis of the Ben Ohau Range, New Zealand

Kirkbride and Matthews (1997) observed the spatial gradient in degree of glacial
3.3 Test applications of the hypsometric analysis

modification of the Ben Ohau Rangen, New Zealand. They also tried to quantify the morphometry and relate it to duration of glaciation. A northward increase in glacial erosion has been established through a morphometric comparison of catchments along the range. In this section, we build on these results and test the possibility of resolving spatial variations in glacial erosion using hypsometry.

Regional setting

The Ben Ohau Range (Figure 3.3) is a 60 km long southern offshoot of the Southern Alps. Ridgeline elevations rise towards the Main Divide, from approximately 1900 m in the south to roughly 2400 m in the north (Kirkbride and Matthews, 1997). Precipitation increases from less than 600 mm yr\(^{-1}\) in the south to over 5000 mm yr\(^{-1}\) in the north (Fitzharris, 1988). The north of the range is characterized by a glacial trough-and-arête morphology, with a temperate glacial climate, whereas the dry south has rounded divides and plateau remnants dissected by fluvial valleys (Kirkbride and Matthews, 1997).

The hypsometric analysis

We use the SRTM data (Farr et al., 2007) to compute the hypsometry of the Ben Ohau Range, New Zealand. As the methodology proposed in this study is more suitable for large scale studies, we first calculate the optimal cell size for the whole Southern Alps, as described in section 3.2.3 (Figure 3.4a). It is found to be 200 by 200 pixels, which is on a geographical WGS projection approximately equal to \(\sim18\) km and \(\sim12\) km in the north-south and east-west directions respectively. The elevations within each cell are binned at elevation intervals of \(\sim30\) m. The altitudes are not corrected for the presence of ice in the landscapes, which is a source of uncertainty. However, at the scale of the analysis, the error can be neglected as the ice cover is relatively small (\(<15\)%). Porter (1975) has defined the lowest LGM ELA of the Southern Alps as approximately equal to 1000 m. This elevation, which is most likely to be lower than the mean long-term ELA, provide a secure limit to separate the fluvially and glacially reshaped domains of the region and calculate \(S_f\) as explained above (section 3.2.3). \(S_f\) is found to be equal to 1.60 ± 0.43 (Table 3.1). For each cell, the hypsometric curve, \(HI\) and \(H_{kr}\)
3. Hypsometric analysis to identify spatially variable glacial erosion

Figure 3.3: Modified from Kirkbride and Matthews (1997); location and simplified geomorphology of the Ben Ohau Range. Morphologic features such as pinnacled, sharp and rounded ridges as well as cirque floors and plateaux are interpreted to indicate a northward increase of glacial imprint. The hypsometry of pinnacled and rounded ridges in a glaciated environment as the one of the Ben Ohau Range are ideally represented by Figure 3.1d and f respectively (the panels on the right indicate their location within the Range).

are computed.

The morphology of the ridges (pinnacled, sharp and round, as defined by Kirkbride and Matthews (1997), Figure 3.3) as well as cirque floors and plateaux enabled us to recognize the northward increase of glacial imprint along the Ben Ohau Range. While this morphologic gradient is seen above 1000 m, post-glacial sediments and glacial lakes hide most part of the bedrock below this elevation, Figure 3.3. Therefore, to compare our analysis with the work of Kirkbride and Matthews (1997) we show, in Figure 3.4b, the hypsometric curves of the cells belonging to the higher domain of the Ben Ohau Range (i.e., at least 50% of the cell area above 1000 m). As predicted in the idealized case, the hypsometric curve is decreased by glacial erosion, with a consequent decrease of $HI$ and increase of
3.3 Test applications of the hypsometric analysis

Figure 3.4: a) Variogram of the hypsometric analysis over the Southern Alps, New Zealand (see text for details); the dashed line represents the selected length of the squared cell (∼18 km N-S and ∼12 km W-E). b) Hypsometric curves of the cells labeled with 1, 2, and 3 in c and d, representing the higher domain of the Ben Ohau Range (at least 50% of the cell area above 1000 m). Light and dark grey represent morphologies characterized by little and extensive glacial erosion respectively. Dotted lines represent the $H_{kr}$ of each hypsometric curve. c-d) Maps of $HI$ and $H_{kr}$ of the Ben Ohau Range, New Zealand. The green lines highlight the main ridge.

$H_{kr}$. Thus, the departure observed between the idealized hypsometric curves proposed in Figures 3.1, 3.2 and 3.3 and the actual hypsometric curves of a natural landscape is relatively small.

The produced maps (Figure 3.4c and 3.4d) show the spatial distributions of $HI$ and $H_{kr}$. Even if the cell size determined by the variogram analysis over the Southern Alps produces a rather coarse subdivision of the Ben Ohau Range, both
3. Hypsometric analysis to identify spatially variable glacial erosion

parameters suggest a northwards increase in glacial erosion which is in agreement with the work of Kirkbride and Matthews (1997). However the increase of $H_{kr}$ is more significant ($\sim 75\%$) than the decrease of $HI$ ($\sim 20\%$), in turn suggesting a higher sensitivity of $H_{kr}$ to glacial erosion.

We also use this test application to investigate how the choice of $S_f$ influences the results. Figure 3.5 shows the maps of $H_{kr}$ of the Ben Ohau Range calculated for one standard deviation away from the mean $S_f$. As expected, the maps show differences as $H_{kr}$ assumes smaller values if $S_f$ is decreased and vice-versa. However, despite the different values, the relative trend identifying the increase of glacial erosion from south to north along the range is clearly visible in all the maps. We conclude that, within the range of reasonable values (see also Table 3.1), the choice of $S_f$ does not significantly influence the final results and interpretation.

![Maps of $H_{kr}$ of the Ben Ohau Range calculated for $S_f - 1\sigma$, $S_f$ and $S_f + 1\sigma$.](image)

Figure 3.5: Maps of $H_{kr}$ of the Ben Ohau Range calculated for $S_f - 1\sigma$, $S_f$ and $S_f + 1\sigma$. Note that, despite the different values of $S_f$, the maps highlight similar trends, consistent with the increase of glacial erosion from south to north along the range.

3.3.2 Numerical Analysis

In this section, we use a numerical model to further explore how fluvial and glacial erosion influence the hypsometry of synthetic and real topographies. We emphasize that the aim of the numerical analysis presented here is to provide information about the first order differences between glacial and fluvial landscapes and their effects on the hypsometry of mountainous regions; we do not aim at fully describing the natural evolution of landscapes using the numerical model.
3.3 Test applications of the hypsometric analysis

The Numerical Model

We use the numerical landscape evolution model ICE-CASCADE (Braun et al., 1999; Tomkin, 2000; Tomkin and Braun, 2002; Herman and Braun, 2008), in which the ice velocities are computed based on the shallow ice approximation (Hutter, 1983) and glacial erosion is simply taken to be a function of the ice-sliding velocity at the ice-bedrock interface (Hallet, 1979, 1996),

$$\frac{\partial z}{\partial t} = K_g |u_s|^l$$

(3.5)

where $u_s$ (m a$^{-1}$) is the sliding velocity of the ice, $K_g$ (m$^{1-l}$ a$^{-l}$) is the glacial erosion constant and $l$ an arbitrary exponent (set equal to 1 in this work). In regions that are not covered by ice, fluvial erosion and hillslope processes are applied. To parameterize fluvial incision, the model adopts a classical power law relationship (e.g., Howard, 1994; Whipple and Tucker, 1999),

$$\frac{\partial z}{\partial t} = K_f A^m (\nabla z)^n$$

(3.6)

where $K_f$ (m$^{(1-2m)}$ a$^{-1}$) is the erosion constant, $A$ (m$^2$) the drainage area and $\nabla z$ is the local slope; $m$ is set to 1/2 and $n$ to 1. When integrated over long timescales hillslope processes are often represented by a linear diffusion equation (Gilbert, 1877) of the form,

$$\frac{\partial z}{\partial t} = D \nabla^2 z$$

(3.7)

where $D$ (m$^2$ a$^{-1}$) is a constant. In the following, the hillslope processes will be referred to as diffusion. The general parameter settings for the computation of the ice velocities are shown in Table 3.2. A more detailed description of the model and the model parameters is provided in Braun et al. (1999); Tomkin (2000); Herman and Braun (2008).

Numerical Simulations

For each simulation, we explore the morphologic changes of the topography by describing the temporal evolution of the hypsometric curve, $HI$ and $H_{kr}$ computed
3. Hypsometric analysis to identify spatially variable glacial erosion

Table 3.2: General parameters settings, refer to Herman and Braun (2008) for more detailed descriptions of the parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$B$</td>
<td>$6.8 \times 10^{-24}$ Pa$^{-3}$ a$^{-1}$</td>
<td>flow law parameter</td>
</tr>
<tr>
<td>$B_s$</td>
<td>$5 \times 10^{-13}$ Pa$^{-3}$a$^{-1}$m$^{-2}$</td>
<td>sliding law parameter</td>
</tr>
<tr>
<td>$p$</td>
<td>3</td>
<td>Glen’s law parameter</td>
</tr>
<tr>
<td>$\rho$</td>
<td>910 kg m$^{-3}$</td>
<td>density of ice</td>
</tr>
<tr>
<td>$g$</td>
<td>9.81 m s$^{-1}$</td>
<td>gravitational acceleration</td>
</tr>
<tr>
<td>$G$</td>
<td>110 mW m$^{-2}$</td>
<td>geothermal gradient</td>
</tr>
<tr>
<td>$K$</td>
<td>$9.828$ W m$^{-1}$K$^{-1}$</td>
<td>thermal conductivity</td>
</tr>
<tr>
<td>$K_d$</td>
<td>$G/\rho 2115$</td>
<td>thermal diffusivity</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>43.5</td>
<td>latitude</td>
</tr>
<tr>
<td>$K_c$</td>
<td>1000km</td>
<td>constriction constant</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>0.006 C m$^{-1}$</td>
<td>lapse rate</td>
</tr>
<tr>
<td>$\gamma_1$</td>
<td>1.</td>
<td>arbitrary constant</td>
</tr>
<tr>
<td>$\gamma_1$</td>
<td>0.2</td>
<td>arbitrary constant</td>
</tr>
</tbody>
</table>

over the whole grid. The elevations are binned using 35 bins whose maximum size is approximately 50 m. In each simulation the rainfall and snowfall patterns are uniform and the sliding and deformation constants are chosen to obtain realistic velocities (i.e., within an order of magnitude of estimates of ice velocities in modern glaciers, $\sim 0–250$ m a$^{-1}$ (e.g., Kaab et al., 1997; Kaab, 2002; Scherler et al., 2008; Herman et al., 2011a)).

Table 3.3: Parameters setting for the simulations 1, 2 and 3.

<table>
<thead>
<tr>
<th>Simulation - 1</th>
<th>Time(ka)</th>
<th>Time Step(ka)</th>
<th>Uplift(mm yr$^{-1}$)</th>
<th>$K_g$(m$^{-1}$ a$^{-1}$)</th>
<th>$D$(m$^2$ a$^{-1}$)</th>
<th>Min-Max Sea Level Temp. (°C)</th>
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<tr>
<td>2300</td>
<td>10</td>
<td>1</td>
<td>40 $\times 10^{-5}$</td>
<td>0.009</td>
<td>2 – 9</td>
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</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Simulation - 2</th>
<th>Time(kyr)</th>
<th>Time Step(kyr)</th>
<th>Uplift(mm yr$^{-1}$)</th>
<th>$K_g$(m$^{-1}$ a$^{-1}$)</th>
<th>$D$(m$^2$ a$^{-1}$)</th>
<th>Min-Max Sea Level Temp. (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>400</td>
<td>1</td>
<td>0</td>
<td>9 $\times 10^{-5}$</td>
<td>0.005</td>
<td>4 – 9</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Simulation - 3</th>
<th>Time(kyr)</th>
<th>Time Step(kyr)</th>
<th>Uplift(myr$^{-1}$)</th>
<th>$K_g$(m$^{-1}$ a$^{-1}$)</th>
<th>$D$(m$^2$ a$^{-1}$)</th>
<th>Min-Max Sea Level Temp. (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>100</td>
<td>1</td>
<td>1 $\times 10^{-4}$</td>
<td>0.001</td>
<td>6 – 13</td>
<td></td>
</tr>
</tbody>
</table>
3.3 Test applications of the hypsometric analysis

Simulation 1 - Transition from Fluvial to Glacial Landscape (Synthetic Topographies)

Figure 3.6: a-d) Four selected snapshots of the simulation showing the uplift and formation of the fluvial topography and, once the steady-state condition is reached, the subsequent glacial cycles. e-h) Four selected snapshots (simultaneous to a-d) showing the spatial distribution of the altitudes variation \( \frac{\partial z}{\partial t} = \text{erosion rate} - \text{uplift rate} \). i) Hypsometric curves of the topographies before (steady-state condition) and after glaciation (light and dark gray respectively). The dashed line highlights the final value of \( H_{kr} \).

The goal of this simulation is to investigate the transition from a theoretical fluvial to a glacial landscape, in a region that is undergoing constant uplift. The model (refer to Figure 3.6, Table 3.3) is applied over a square region of approximately 30 by 30 km and runs for 2.3 Myr. The initial topography is flat and is uplifted at a constant rate of 1 mm yr\(^{-1}\). The model runs until a fluvial steady-state is reached, after which glacial erosion is introduced. Three glacial cycles are
applied by lowering the ELA, in turn glaciating the mountain range and reshaping the fluvial morphology. The period of oscillation of the ELA is 100 kyr and shifts from an altitude of $\sim 1400$ m down to $\sim 100$ m (it therefore represents a relatively extreme case). The model does not allow cold-based glaciers. $S_f$ for this simulation, deduced from the hypsometric curve of the whole grid at the fluvial steady-state condition, is found to be equal to 1.80. To fully isolate the effects of fluvial and glacial erosion, the model does not compute isostatic compensation in response to the erosion and diffusion is minimized.

Figure 3.6a-h shows the evolution of the model run. The steady-state fluvial topography can be compared to that of a small range. The use of a single hypsometric curve over the whole grid does not allow the identification of glacially induced morphologic gradients within the modeled range. Nonetheless, this choice enables us to characterize the glacially induced decrease of the hypsometric curve (Figure 3.6i). The significant departure between the pre- and post-glaciation values of both $HI$ and $H_{kr}$ suggests the sensitivity of the morphologic parameters to glacial reshaping. Finally, the growth of $H_{kr}$ recorded over the whole glacial period is more significant than the decrease of $HI$, suggesting a higher sensitivity of the former to glacial erosion.

Simulation 2 - Transition from Fluvial to Glacial Landscape - Alpi Apuane Domain (Apennines), Italy

We now apply the landscape evolution model to real topography. A very limited glacial imprint is recorded in the Alpi Apuane domain of the Apennines, Italy (Giraudi, 2004) (see also section 4.1). It can, therefore, be considered as a suitable region to study the transition from fluvial to glacial landscape. This simulation (see Figure 3.7 and Table 3.3 for further details about parameters settings) is applied over a rectangular region of approximately 8 by 11.5 km.

The model is run for 400 kyr, during which time four glacial cycles are applied (Figure 3.7a-h). To simulate pervasive glaciations, the ELA is decreased from about 1500 m down to approximately 500 m. Rock uplift is minimized to limit the influence of tectonics on the temporal evolution of $HI$ and $H_{kr}$. Diffusion is neglected and only minimal fluvial erosion is allowed. The model does not allow cold-based glaciers. Given the fluvial nature of the Alpi Apuane domain $S_f$ is calculated from the hypsometric curve of the current morphology over the whole
3.4 Application to the European Alps and the Northern Apennines

The different glacial experience of the European Alps and the northern Apennines is certainly one of the causes of their sensibly different morphologies. The appli-
3. Hypsometric analysis to identify spatially variable glacial erosion

Figure 3.7: a-d) Four selected snapshots of the simulation showing the occurrence of the glacial cycles. e-h) Four selected snapshots (simultaneous to a-d) showing the spatial distribution of the altitudes variation ($\frac{\partial z}{\partial t} = \text{erosion rate} - \text{uplift rate}$). i) Hypsometric curves of the topographies before and after glaciation (light and dark grey respectively). The dashed lines highlight the values of $H_{kr}$.

cation of the newly-developed methodology to this particular setting enables us to further test its efficacy and constrain its use as well as gaining more information about the role of glacial erosion in shaping mountain ranges.

3.4.1 Regional Setting

Hanging valleys, glacial cirques, “U”-shaped valleys, moraines and glacially polished rocks demonstrate the importance of past glaciations in the European Alps
3.4 Application to the European Alps and the Northern Apennines

Figure 3.8: a-d) Four selected snapshots of the simulation showing the evolution of the Zermatt-Saas catchment undergoing fluvial erosion. e) Hypsometric curves of the initial and final topographies (dark and light gray respectively). The dashed lines highlight the position of $H_{kr}$ of the initial and final stages of the simulation.

(e.g., Ivy-Ochs et al., 2006b). The impact of Quaternary Glaciations on the topography of the Alps, however, is still under debate: during the LGM, glaciers expanded onto the foreland after 30 kyr (Ivy-Ochs et al., 2006a) and long-term denudation rates determined by low-temperature thermochronology are in general in the range of 0.2 to 1 mm yr$^{-1}$ (e.g., Rahn, 2001; Fugenschuh and Schmid, 2003; Malusa et al., 2005; Luth and Willingshofer, 2008; Schlunegger and Willett, 1999).

Systems of moraines in the Alpine valleys record repeated glacial advances and retreats (e.g., Heuberger, 1966; Gross et al., 1978; Maisch, 1981, 1987; Ivy-Ochs et al., 2009) caused by a cyclic shift of the ELA within a range of about 1000 m, from 2900 m a.s.l. (the present-day ELA) to 1800±100 m a.s.l, during the upper Quaternary (e.g., van der Beek and Bourbon, 2008).

In comparison, evidence of glaciation in the Apennines remains sporadic (Giraud, 2004). The systems of moraines of this region have been correlated with those of the Alps and the oldest moraines are related to a generic pre-Würmian event (Jaurand, 1994). The upper Pleistocene LGM glaciers started to expand earlier than 22.6±0.63 kyr, and began to retreat around 21.5 kyr (Giraudi, 2004). At present, only a single small glacier that is rapidly melting is found in the
3. Hypsometric analysis to identify spatially variable glacial erosion

Apennines: the Calderone Glacier in the Gran Sasso Massive, which is the most southerly glacier in Europe. Beside these small glacial features, fluvial erosion and hillslope processes have clearly dominated.

The tectonics of the whole study region is dominated by counter-clockwise rotation of the Adriatic plate with respect to stable Eurasia, around a pole located in the Po plain (Anderson and Jackson, 1987; Calais et al., 2002; D’agostino et al., 2008). This rotation controls the extension across the Apennines and the western Alps as well as the shortening in the central and eastern Alps. Although the origin of the current deformation of the Alps is still unclear, geodetic data show a rock uplift in the Swiss Alps of 1-2 mm yr$^{-1}$ (Kahle et al., 1997). At present, GPS data show limited (less than 2 mm yr$^{-1}$) east-west extension in the Western Alps (Calais et al., 2002; Sue et al., 2007). The lack of present day convergence in the Western Alps, together with the observation of sediment-sealed thrusts in the western part of the Po basin (Pieri and Groppi, 1981), and the cessation of thin-skinned deformation in the Jura Mountains at ca. 4 Myr (e.g., Becker, 2000) suggest very limited current orogenic activity within the range.

3.4.2 Hypsometric Analysis

We use the SRTM data (Farr et al., 2007) to compute the hypsometry of the study region (Figure 3.9a) and a variogram is used to define the optimal cell size (Figure 3.9b). It is found to be 300 by 300 pixels cell, which is on a geographical WGS projection approximately equal to $\sim$27 km and $\sim$18 km in the north-south and west-east direction, respectively. The elevations within each cell are binned at elevations intervals of $\sim$30 m. The altitudes are not corrected for the presence of ice in the landscapes. However, because of the large scale of the analysis, we believe that the error can be neglected as the current ice cover is a small percentage of the study region (< 10%). The LGM ELAs of the study region vary between approximately 1000 m for the Alps, and 1700 m for the Appennines (Benz, 2003; Buoncristiani and Campy, 2004; Giraudi, 2004). $S_f$ is therefore obtained by averaging $\bar{S}$ for the hypsometric curves of the cells whose maximum elevation is below 1000 m. $S_f$ is found to be equal to 1.82±0.54. For each cell the hypsometric curve, $HI$ and $H_{kr}$ are computed (Figures 3.9c-d).
Most peripheral valley bottoms are characterized by very low slopes, due to post-glacial sediment infill (van der Beek and Bourbon, 2008), that may affect the morphometric analysis. For this reason we also show in Figure 3.9c-d, the limits of the post-glacial sediment infill of the Alpine domain. These limits, that roughly follow the 600 m isoline, were traced with respect to the grid of squared cells used for the morphometric analysis. They separate the axial massifs of the Alps, where the sediments are produced, from the peripheral domains of the chain where post-glacial sediments are more likely to deposit, filling the original morphology of the valley bottom.

In the map of $HI$ (Figure 3.9c), the Alpine arc is distinguished by a cluster of cells with values around 0.5. More importantly, no significant trends are recognizable within the Alpine belt. Cells with similar values, however, can be seen outside the Alps (e.g., in the Po plain, as well as in the Molasse basin, in Sardinia and Corsica). The highest values are located in the Alpine piedmont region and in the Apennines, while the lowest values, without considering the cells along the coastline, are seen in the Po delta region. On the contrary, the map of $H_{kr}$ (Figure 3.9d) shows the Alps to be the only zone with high values, highlighting the presence of glacial morphologies. Within the Alps, furthermore, it is possible to identify several and wide variations in the value of $H_{kr}$ which might represent, as for the test example of the Ben Ohau Range, different degrees of glacial imprint within the mountain belt. The regions covered by post-glacial sediments, in general, show high $HI$ and low $H_{kr}$, which was also observed in the values obtained for the Ben Ohau Range.

### 3.5 Discussion

Previous studies have used hypsometry to characterize fluvial and glacial landscapes (e.g., Strahler, 1952; Brockethurst and Whipple, 2004). Although its use has been demonstrated, it remains difficult to understand the relative importance that glacial and fluvial processes have on hypsometry. We focus here on the effects that glacial erosion processes have on the morphologic evolution of a valley. Even though glacial and/or peri-glacial processes are complex, we started from the simplified assumption that first order glacial morphologies can be represented
3. Hypsometric analysis to identify spatially variable glacial erosion

Figure 3.9: a) Region included in the hypsometric analysis. The white line represents the last glacial maximum (LGM) ice extent and the yellow lines delimit the region where glaciated bedrock is exposed (see text for further details). b) Variogram of the hypsometric analysis (see text for details); the dashed line represents the selected length of the squared cell (∼27 km N-S and ∼18 km W-E). c-d) Maps of $HI$ and $H_{kr}$. The white line represents the LGM ice extent (Ehlers and Gibbard, 2004). The green lines delimit the region where glaciated bedrock is exposed (see text for further details).

by rounded valley bottoms. Local factors preventing bed erosion (e.g., cold-based glaciers, subglacial hydrology or sediment protection) will obviously affect the interpretation arising from the methodology developed here, but ice sliding velocity patterns at the glacier-bedrock interface will often lead to a “U”-shaped valley (Harbor et al., 1988; Harbor, 1992). It is shown here that such a change on cross-
sectional valley profiles results in a localized decrease of the hypsometric curve, as previously suggested (Brocklehurst and Whipple, 2004). Consequently, in addition to the insight provided by $HI$, the gradient of the hypsometric curve can give information about the glacial signature on a landscape. We exploit this characteristic to define $H_{kr}$, as the minimum normalized elevation at which the gradient of the hypsometric curve equals $S_f$. We test the hypsometric analysis on a reasonably well-known region, the Ben Ohau Range, and a series of numerical models.

The spatial patterns of $HI$ and $H_{kr}$ values suggest a higher capability of $H_{kr}$ to highlight regions were glacial erosion was important. Furthermore the produced hypsometric maps can provide useful information about the spatially variable glacial imprint, as illustrated using the present-day topography of the Ben Ohau Range. The results clearly show how the observed latitudinal gradient in glacial imprint (Kirkbride and Matthews, 1997) can be highlighted by both hypsometric parameters. It is, however, clear that factors such as post-glacial sediment in-fill of overdeepenings strongly limit the applicability of the hypsometric analysis, as they might confer a sub-horizontal geometry to the valley bottom that affects the hypsometric curve. Sub-horizontal sediment in-fill induces a decrease in the hypsometric curve gradient near the origin followed by an abrupt increase where the steeper bedrock of the valley flanks is exposed. In general, this leads to a low $H_{kr}$ value as the gradient of the hypsometric curve gets higher than $S_f$ when the normalized elevation has grown relatively little. This is, for instance, the case of the lower part (i.e., below 1000 m) of the Ben Ohau Range, where a low $H_{kr}$ is obtained.

Another important limitation could come from the fact that the hypsometric curve is normalized by the elevation range, making the behavior of the ridgelines critical. Consider a simple fluvial landscape with a fixed base-level in which a few modest high altitude cirques are carved. If the cirques are accompanied by ridgeline lowering, the elevation range decreases, and the hypsometric curve is everywhere steeper than the fluvial case (i.e., low $H_{kr}$), except for the details in the cirques, even though the majority of the landscape is unmodified. However, if cirques are formed with no change to ridgeline elevations, the hypsometric curves will be identical up to the elevations where the cirques modify things (i.e., high $H_{kr}$). This could be extended to a tectonically active landscape where cold-based
3. Hypsometric analysis to identify spatially variable glacial erosion

Figure 3.10: HI (a) and $H_{kr}$ (b) versus mean elevation. The LGM ELAs for the considered region are represented in the plot by the grey shaded area. The diamonds represent the average and the error bars are one standard deviation (68.2% confidence) of HI and Hkr for the cells included in the region of study, binned by mean elevation. In the boxes is shown the frequency distribution of the raw data for a selected bin. a) A general decrease of HI with increasing mean elevation is recorded and confirmed by the high absolute value of the Spearman rank-order correlation coefficient ($r_s$) for the entire dataset. b) A strong increase of $H_{kr}$ within the range of the LGM ELAs for the considered region is recorded and confirmed by the high absolute value of the Spearman rank-order correlation coefficient ($r_s$) of the block of elevations including the LGM ELAs. It must be kept in mind that, for elevations below 600 m, sediment accretion significantly affects the values of HI and $H_{kr}$.

ice actually decreases ridgeline erosion compared with the fluvial case. Here $H_{kr}$ can be different for a minor change in circumstances, which must be kept in mind when looking at the results.

The numerical analysis provide a means to assess the sensitivity of the method in identifying switches between fluvial and glacial conditions. The significantly
3.5 Discussion

different values of both $HI$ and $H_{kr}$ before and after glaciation (Simulations 1 and 2) or return to fluvial conditions (Simulation 3), highlight the efficacy of the method. The high variation of $H_{kr}$ confirms that the gradient of the hypsometric curve might complement the information about glacial and/or fluvial erosion provided by $HI$.

The European Alps and Apennines have a different geological and climatic history. The comparison between the results in the Alps and the northern Apennines provides information about the importance that glacial erosion has had on the Alps topography. The map of $HI$ (Figure 3.9c) shows similar values within the Alps, the Po plain, the Molasse basin, Sardinia and Corsica. We interpret this inconsistency as a natural example of the theoretical case illustrated in Figure 3.1a and 3.1f, where different hypsometric curves have similar $HI$. In addition, the cells pertaining to the Alps show a very limited variability of $HI$, in turn suggesting either the presence of an homogeneous morphology, or the inadequacy of using $HI$ alone. On the contrary, the highest values of $H_{kr}$ (Figure 3.9d) are concentrated within the Alpine domain. In this region, $H_{kr}$ shows large variations that we interpret to be indicative of a spatially variable pattern of glacial erosion, similar to the one observed in the test application on the Ben Ohau Range. However, it is worth noting that the strong limitations discussed above are potential sources of bias. The limits that separate the erosional and depositional domains in the Alps are shown in Figure 3.9c and 3.9d to assess where such a bias may arise.

We now compare the distribution of $HI$ and $H_{kr}$ with observations of LGM ELAs (Benz, 2003; Buoncristiani and Campy, 2004; Giraudi, 2004). In Figure 4.6, we show the variation of $HI$ and $H_{kr}$ as a function of the mean elevation of each cell. The analysis highlights an approximately constant decrease of the $HI$ with increasing mean elevation and a strong increase of the $H_{kr}$ corresponding to the LGM ELAs. These trends are also confirmed by the Spearman rank-order correlation coefficients (shown in Figure 4.6), computed for the whole set of raw data and for three different blocks including the elevations below, at and above the LGM ELAs respectively. As suggested by several studies (e.g., Brocklehurst and Whipple, 2004; Anderson et al., 2006; Egholm et al., 2009), the integrated glacial erosion over glacial periods is maximized around and above the mean long term ELA. The relationship between $H_{kr}$ and mean elevation further indicates that
the morphometric parameter is more sensitive to changes in the amount of glacial erosion than to variations in lithology or tectonics as the study area encompasses several lithological sequences and tectonic units. The correlation between increased glacial erosion and LGM ELAs has been interpreted, on other mountain ranges, as the prevalence of the glacial buzz-saw (e.g., Brozovic et al., 1997; Brocklehurst and Whipple, 2004) thus indicating that climate has some control on the large-scale landscape development. The analysis proposed here hints that this could also be the case of the European Alps, as recently suggested by Anders et al. (2010). It is worth pointing out that, despite the increase of $H_{kr}$ corresponding to the LGM ELAs of the European Alps, the highest glacial signature (recognized by the highest values of $H_{kr}$) is found at slightly higher elevations (between 1500 m and 2100 m, Figure 4.6b). A possible interpretation is the correlation of the highest glacial signature determined by the distribution of $H_{kr}$ with the Most Extensive Glaciation (MEG) ELA which has been estimated by van der Beek and Bourbon (2008) as 1800±100 m in the western Alps. This interpretation must, however, be treated with caution as large overdeepenings are observed in the Alps, which cannot be assessed using the SRTM data alone.

3.6 Conclusions

We have explored the effects that glacial erosion has on the hypsometric curve of idealized cross-sectional valley profiles. This approach highlights that very different valley geometries can lead to similar $HI$. The same analysis suggests that the gradients of the hypsometric curve may provide less ambiguous insights into the amount of glacial erosion experienced by a landscape.

We propose a way to resolve the glacially induced modifications on hypsometry by defining a new morphometric parameter, called the hypokyrtome, $H_{kr}$, as the minimum normalized elevation at which the gradient of the hypsometric curve is greater or equal to a reference value ($S_f$). We test its applicability on the Ben Ohau Range, New Zealand, which shows that hypsometry is able to describe spatially variable glacial imprints and underlines the high sensitivity of $H_{kr}$. We then perform a numerical analysis that helps to further constrain the use of $H_{kr}$.

We apply the hypsometric analysis to a region which includes the European
3.6 Conclusions

Alps and the northern Apennines. The hypsometric parameters provide insights into the major regional morphologic gradients induced by glaciation. The correspondence between the concentration of surface area and the LGM ELAs suggest the prevalence of a glacial buzz-saw.

Even if the study of spatially variable glacial imprint on mountain ranges through hypsometric analysis can help recognize whether glacial activity has been significant, the modifications induced by tectonics, lithology, hillslope processes and/or precipitation rates on hypsometry, are still largely unknown. Thus, it must be kept in mind that local variations of such factors may affect the analysis.
Chapter 4

Pre-glacial topography of the European Alps


4.1 Introduction

Spectacular erosional landforms in the European Alps and other mid- to high-latitude mountain belts highlight the importance of glacial erosion in shaping mountain topography (e.g., Penck, 1905). The onset of significant glacial erosion in the Alps dates back to at least \( \sim 0.9 \) Ma (Muttoni et al., 2003) and is thought to have enhanced exhumation and rock uplift by increasing average erosion rates (Champagnac et al., 2008), accompanying an increase of topographic relief (e.g., Haeuselmann et al., 2007; Norton et al., 2010; Valla et al., 2011). Recent morphometric analyses have also suggested that glacial erosion was maximized around the mean equilibrium line altitude (ELA) (Sternai et al., 2011) or even higher (Anders et al., 2010). Glacial reconstructions and their relations with past climate are relatively well established in the Alps (e.g., Florineth and Schlüchter, 2000; Kelly et al., 2004), but at present there is no simple methodology for reversing glacial erosion and estimating the Alpine topography prior to glaciation.
An estimate of the pre-glacial topography would be valuable for two purposes. First, the difference between pre-glacial and modern topography provides an estimate of the magnitude and spatial distribution of glacial erosion, thereby constraining glacial erosional processes. Second, removing the glacial overprint could reveal important differences in regional topography that reflect tectonic processes and precipitation patterns prior to glaciation. Some reconstructions have been proposed for pre-glacial topography of the Alps at the local (Glotzbach et al., 2011; Haeuselmann et al., 2007; Valla et al., 2011) or regional scale (Champagnac et al., 2007; Szekely, 2003). These latter studies, however, are based on geometrical estimates of “missing volumes” that have supposedly been eroded during the Quaternary, without any rigorous way of generating a realistic pre-glacial topography.

We present a new method for estimating the topography of the entire Alps based on fundamental geomorphic characteristics of fluvial networks. We assume that glacial erosion has deepened and broadened valleys, but has not greatly eroded river channel heads, an assumption supported by recent thermochronological studies (Glotzbach et al., 2011; Valla et al., 2011, 2012). In addition, we assume that glaciation has not changed the spatial pattern of the river network, so that the relationships between drainage area, and river length are still preserved by the drainage network. With these assumptions, we can reconstruct the pre-glacial topography by connecting the channel heads to base level, following a profile consistent with steady-state fluvial incision and the drainage network geometry. This approach is robust because the branching network structure of river basins imply that each major river is supported by multiple channel heads, thereby providing multiple data constraints on channel steepness. In fact, as we demonstrate in this paper, we can resolve channel steepness that is variable in space, thereby providing information on the pre-glacial rock uplift, climate and erosion patterns as well as the distribution and magnitude of glacial erosion and associated isostatic adjustment.
4. Pre-glacial topography of the European Alps

4.2 Methodology

4.2.1 Fluvial erosion

We assume that the topography of a non-glaciated mountain range is primarily controlled by the balance between rock uplift and fluvial erosion. To model this balance, we use the generalized stream power law (e.g., Howard et al., 1994; Whipple and Tucker, 1999):

$$ \frac{\partial z}{\partial t} = U - KA^m S^n $$

(4.1)

where $z$ is the elevation, $U$ is the rock uplift rate, $A$ is the upstream drainage area, $S$ is the bedrock channel gradient, and $K$ is a dimensional coefficient of erosion (e.g., Whipple et al., 1999). The exponents $m$ and $n$ may be determined theoretically or empirically and typically vary between 0.3 and 1 and between 0.7 and 1, respectively (Tucker and Salingerland, 1994; Whipple and Tucker, 1999). Over Myr timescales, erosion rates are adjusted to those of rock uplift to reach topographic steady-state and equation 4.1 becomes Flints law (Flint, 1974):

$$ S = k_s A^{-\theta} \text{ with } k_s = \left( \frac{U}{K} \right)^{\frac{1}{n}}, \quad \theta = -\frac{m}{n} $$

(4.2)

where $k_s$ is labeled the “steepness index” and $\theta$, referred to as the “concavity index”, is nominally 0.5 for bedrock channels (e.g., Whipple and Tucker, 1999). This relationship has been exploited to resolve spatial variations of present-day $k_s$ patterns associated with changes in rock uplift rate or rock erodibility across tectonic or lithological boundaries, respectively (Norton et al., 2010; Whipple et al., 1999; Whipple and Tucker, 1999; Wobus et al., 2006). Here, we use equation 4.2 to infer spatial variations of $k_s$ that describe the theoretical steady-state morphology of the European Alps prior to glaciation.

4.2.2 Data and analysis

We assume that glacial erosion enhanced lowering of Alpine valleys without significantly modifying channel head elevations, except for a component of isostatic rock uplift in response to glacial erosion for which we explicitly account. With this
assumption, modern elevations of the channel heads, defined to be channels with a contributing area of $\sim 0.4 \text{ km}^2$ (Figure 4.1), become the key constraint in our landscape reconstruction and we can reconstruct the elevation of the full pre-glacial fluvial network by imposing a constant channel concavity, $\theta$, and estimating the channel steepness throughout the network.

Figure 4.1: Location of the base level (white line) and channel head nodes (black dots). The onset shows the Zermat-Saas catchment and surrounding areas during the LGM as mapped by Schlüchter et al. (2009) (reproduced by permission of swisstopo-JA100120) and the channel head nodes matching crest line and nunataks.

The algorithm (see also Figure 4.2) consists of two steps: (1) an initial search for a spatially uniform $k_s$ that minimizes the misfit between predicted and observed channel head elevations; and (2) iterative local perturbation of this initially uniform $k_s$, with a subsequent assessment as to the improvement in channel head elevation prediction. We identified 54500 channel heads, so in principle, we could resolve $k_s$ on 54500 channel segments, but in practice, we expect that $k_s$ varies smoothly over distances of ones to tens of kilometers given that it primarily depends on rock uplift, precipitation rate and rock type (i.e., the ensemble of all rock-related characteristics). Thus, we balance the resolution of $k_s$ in individual
channels against the need to smooth elevation errors in the channel head data, by imposing a spatial smoothness on \( k_s \). We randomly select a position in the topographic domain which serves as the center of a Gaussian perturbation function, \( P_{(x,y)} \), with a specified length scale parameter, \( \sigma \):

\[
P_{(x,y)} = p \exp \left( -\frac{(x - x_0)^2 + (y - y_0)^2}{\sigma^2} \right)
\]  
(4.3)

where \( p \) is the magnitude of the perturbation (also selected randomly), \( x_0 \) and \( y_0 \) are the coordinates of the center of the perturbation and \( x \) and \( y \) are the
coordinates of all nodes within the grid. This process yields an estimate of the pre-glacial channel steepness that is variable but smooth in space.

4.2.3 Isostatic response to Quaternary erosion

Following the onset of glaciation, there is an additional component of rock uplift given by the isostatic re-equilibration in response to glacial erosion. To determine this extra component of rock uplift, we calculate the difference between the reconstructed fluvial topography (derived from the optimization scheme described earlier) and the present-day topography to estimate the volume of rock eroded by Quaternary glaciation. We then use a two-dimensional flexural model to compute the vertical deflection of an elastic plate produced by removal of this material (e.g., Watts, 2001), reduce the present-day channel head elevation by this amount and repeat the data inversion. We iterate this procedure until the average isostatic correction is less than a specified threshold representing a trade-off between quality of the estimates and computational time (i.e., 10% of the value obtained in the first iteration).

4.3 Test applications

4.3.1 King Range, Northern California

Based on the record from marine terraces, the King Range, Northern California, can be divided into high and low uplift regions (Merritts and Bull, 1989). Focusing on the Gitchell, Shipman and Big catchments in the high uplift region and on the Hardy, Howard and Juan catchments in the low uplift region, Wobus et al. (2006) have shown that the topography of the King Range is adequately described by Flint’s law: $S = k_s A^{-\theta}$. Using slope-area data from these catchments, they have estimated an $A_{cr}$ of about $10^5$ m$^2$ and calculated best-fitting mean $k_s$ of 117 m$^{0.9}$ and 66 m$^{0.9}$ for the high and low uplift regions, respectively. We build on these results to test whether our algorithm is able to provide similar results.

We use the 3 arcsec resolution SRTM data (Farr et al., 2007) and focus the study on two squared regions including the Gitchell, Shipman and Big catchments
4. Pre-glacial topography of the European Alps

![Residual misfit against randomly selected, spatially uniform $k_s$. Black and gray circles represent model run on the high and low uplift region, respectively. The best fitting $k_s$ values are equal to $\sim 120\,\text{m}^{0.9}$ for the high uplift region and $\sim 60\,\text{m}^{0.9}$ for the low uplift zone, similar to the ones found by Wobus et al. (2006).](image)

and the Hardy, Howard and Juan catchments, respectively (lower-left and upper-right corners of the first (high uplift) region: 124.219°W-40.101°N; 124.100°W-40.187°N; lower-left and upper-right corners of the second (low uplift) region: 123.819°W-39.661°N; 123.701°W-39.739°N). The present-day coastline provides a natural base level for both areas. We set a reference $\theta$ of 0.45 (the same value used by Wobus et al. (2006)) and calculate the spatially uniform $k_s$ that minimizes the residual misfit between the present-day and modeled topography, for the nodes whose drainage area is $\geq 10^5\,\text{m}^2$. As expected, the results are highly similar to those obtained by Wobus et al. (2006) (Figure 4.3).

### 4.3.2 New Zealand, Southern Alps

The South Island of New Zealand is also a suitable location to test our algorithm. It exhibits a strong contrast in rock uplift rates, exhumation rates and lithology along the plate boundary between the Australian and Pacific plates (namely the Alpine Fault), but a rather uniform climate (Figure 4.4). In the central part, i.e., the Southern Alps, the lithology mainly consists of easily erodible greywacke and schists (e.g., Cox and Sutherland, 2007). Low-temperature thermochronological data suggest exhumation rates adjacent to the Alpine Fault between 6 and 8 mm yr$^{-1}$ (Figure 4.4e, Tippett and Kamp, 1993b,a; Batt et al., 2000; Herman et al., 2007, 2009, 2010), while landslide analysis (Clarke and Burbank, 2010; Hov-
4.3 Test applications

Figure 4.4: Maps of the South Island of New Zealand showing (a) digital terrain model of topography; (b) mean annual rainfall, modeled for the period 1971-2000 (Tait et al., 2006); (c) glacial geology and preserved Cretaceous-Tertiary erosion surface (Cox and Sutherland, 2007) with extent of ice at the last glacial maximum (white) (Barrell, 2011) and active faults with known evidence for rupture in the past 120,000 years (GNS Science, Active faults database 2011, from http://data.gns.cri.nz/af/); (d) erosion, calculated as a mean ground lowering from a suspended sediment yield model (Hicks et al., 1996), assuming an average crustal density of 2.65 t/m$^3$; (e) thermochronologic data; (f) maximum rates of contemporary shear strain derived by GPS surveys between 1996 and 2008 (updated from Beavan et al. (2007)) together with Pacific-Australian plate vectors calculated at Fox Glacier (Beavan et al., 2007; DeMets et al., 2010) and central Alpine Fault late Quaternary slip rates (Norris and Cooper, 2001); and (g) reconstructed pattern and magnitude of $k_s$.

ius et al., 1997) and annual sediment yield estimates (Griffiths, 1981; Hicks et al., 1990, 1996) reveal erosion rates of up to 9 mm yr$^{-1}$ (Figure 4.4b). These measurements suggest steady-state conditions between rock uplift and erosion rates (Adams, 1980; Herman et al., 2010). To the south, the Fiordland region exhibits lower rates of exhumation, around 1-2 mm yr$^{-1}$ (Shuster et al., 2011). In contrast to the Southern Alps the, lithology of the Fiordland region mainly consists of less erodible granites.

We use the 3 arcsec resolution SRTM data (Farr et al., 2007) over a region
that includes the entire South Island. The present-day coastline provides a natural base level for both areas. We set a reference $\theta$ of 0.5 (an arbitrary value within the variability range of theoretical fluvial landscapes (Whipple and Tucker, 1999)) and calculate the spatially variable $k_s$ as described above.

Results highlight a strong spatial correlation between exhumation rates (Figure 4.4e) and $k_s$ (Figure 4.4g). $k_s$ values in the central Southern Alps are greater than in Fiordland by approximately a factor of 1.5 (Figure 4.4g). Rock uplift can also be estimated directly from exhumation rates (Figure 4.4e), given the steady-state conditions. We can thus write:

$$\frac{k_s(SA)}{k_s(F)} = \frac{(U/K)(SA)}{(U/K)(F)} \approx 1.5 \quad (4.4)$$

where the indices SA and F stand for “central Southern Alps” and “Fiordland”, respectively. Given that $U_{SA} \approx 3U_{F}$ (Figure 4.4e), we then obtain:

$$K(SA) \approx 2K(F) \quad (4.5)$$

indicating that the granites of the Fiordland region are roughly two times less erodible than the greywacke and schists of the central Southern Alps, in turn affecting the channel the steepness and landscape morphology. This test, therefore, demonstrates the robustness of the newly-developed methodology.

## 4.4 Reconstruction of the pre-glacial $k_s$ and Alpine topography

### 4.4.1 Experiment setting

We used re-sampled SRTM data (Farr et al., 2007) to calculate the pre-glacial topography of the European Alps. The best trade-off between resolution and computational time was found for horizontal resolutions of $\sim 540$ m and $\sim 780$ m in the longitudinal and latitudinal directions, respectively. Concavity indices for glacial and non-glacial catchments on the present-day topography are statistically indistinguishable, averaging 0.73 and 0.71, respectively (Norton et al., 2010). We
expect pre-glacial concavity to be lower than the modern value and set the parameters \( m \) and \( n \) equal to 0.5 and 1, respectively (i.e. reference \( \theta \) equal to 0.5), also assuming that pre-glacial fluvial erosion occurred mainly on bedrock channels (Whipple and Tucker, 1999). Hillslope processes limit channel gradients upstream of a critical drainage area, \( A_{cr} \), and Flints law is only valid below this point (Stock and Dietrich, 2003). 70% to 90% of topographic relief, however, is represented by fluvial channels where Flints law is valid (Whipple, 2004), and when the spatial resolution is in the order of hundreds of meters, as the case of this work, debris flow and channel nodes are indistinguishable (e.g., Beaumont et al., 1992; Chase, 1992). The base level (Figure 4.1 and 4.5) is traced along the perimeter of the mountain range and follows the boundaries between the Alpine tectonic units and the peri-alpine sedimentary cover. Where this boundary cannot be identified (e.g., between the Alpine-Dinarids and Alpine-Apenninic domains) the base level was simply defined to form a closed polygon following, as far as possible, the boundaries between other tectonic units (e.g., Schmid et al., 2004, for a reference tectonic map of the Alpine orogen). The ordering of the nodes is based on the CASCADE algorithm (Braun and Sambridge, 1997), and the top-of-the-stack nodes were chosen as channel head nodes (Figure 4.1). We only select channel heads lying above the mean topography (calculated through a \( \sim \) 20x20 km sliding window) plus 1\( \sigma \) standard deviation, to avoid taking into account high frequency peak elevations and associated channel head nodes that are most likely to be eroded by glacial activity (Figure 4.5). The selected channel heads (a total of 54500 channel head nodes over 439353 nodes in the considered grid) match closely with crest lines and reconstructions of Last Glacial Maximum (LGM) nunataks (Kelly et al., 2004; Schlüchter et al., 2009), which gives support to the assumption that glacial erosion did not impact these elevations (Figures 4.1). The isostatically driven rock uplift induced by enhanced erosion during the glacial period was calculated using spatially uniform crust and mantle densities (2.7 and 3.3, respectively). Poissons ratio was set to 0.25 (Christensen, 1996) and the elastic thickness of the crust was set to 20 km (Stewart and Watts, 1997).
4. Pre-glacial topography of the European Alps

Figure 4.5: Selection of the channel head nodes. The present-day topography, the profile locations (white straight lines) and the base level (white curve line) are shown in the map. In the profiles, the solid and dashed lines represent the present-day and mean topographies, respectively. Dotted lines are one standard deviation away from the mean topography. Gray circles represent the selected channel head nodes. The mean topography is calculated using a \( \sim 20 \) by 20 km sliding window.

4.4.2 Choice of the preferred model

Several random seeds were tested to ensure that the final solution is not dependent on the intrinsic function of Fortran 90 used to generate pseudorandom numbers.

The residual misfit between present-day and modeled channel head elevations increases with increased length scale (Figure 4.6a). When the length scale is \(<\sim 30\) km the residual misfit drops rapidly indicating that the solution is overdamped. On the other hand, a length scale \(>\sim 30\) km do not allow to significantly reduce the residual misfit. The shape of the curve, however, confirms that the model results contain coherent information and the preferred model should be chosen near the corner of the curve (e.g. Soldati et al. 2006). We chose \(\sim 30\) km (\(\sim 60\%\) of the residual misfit at a length scale of \(\sim 200\) km, a distance similar to the average width (N-S) of the mountain range) as optimal length scale for \(k_s\) in the Alps. The results of the experiment with a length scale of \(\sim 30\) km are shown in Figure 4.6b-c.

To establish how well each modeled parameter (i.e., pre-glacial \(k_s\) values) is resolved with respect to data points (i.e., channel head nodes), we show in Figure 4.7 the normalized standard deviation (1\(\sigma\)) of the variation of residual misfit produced by a number of modulations (within \(\pm 100\) m) of the calculated pre-glacial
4.4 Reconstruction of the pre-glacial $k_s$ and Alpine topography

Figure 4.6: a) The residual misfit of different experiments is plotted against length scales. b) First part of the calculation procedure for the experiment with a length scale of approximately 30 km, where the spatially uniform $k_s$ minimizing the residual misfit between modeled and present-day channel head elevations is sought. c) Second part of the calculation procedure for the experiment with a length scale of 30 km where, at each iteration, the best fitting spatially uniform $k_s$ is perturbed to further reduce the residual misfit at the channel heads. NN is the number of nodes in the considered grid.

$k_s$ at each node.

Finally, we explored the sensitivity of the reconstructed pre-glacial $k_s$ to potential lowering of channel head elevations by glacial erosion and reference concavity.
4. Pre-glacial topography of the European Alps

index, \( \theta \) (Figures 4.8 and 4.9, respectively).

Figure 4.8: Reconstructed pre-glacial \( k_s \) map obtained assuming an arbitrary glacially-induced decrease of channel head elevations equal to 25% (a) and 50% (b) of their heights on the present-day topography. Other settings are the same as those of the preferred model. The magnitude of pre-glacial \( k_s \) appears to be sensitive to potential lowering of the channel heads by glacial erosion, but the pattern does not (see also Figure 4.10b).

Figure 4.9: Reconstructed pre-glacial \( k_s \) with a reference concavity index, \( \theta \), equal to 0.4 (a) and 0.7 (b). Other settings are the same as those of the preferred model. While the magnitude appears to be highly sensitive to the reference concavity index, the overall pattern remain similar.

4.4.3 The reconstructed pre-glacial \( k_s \) and alpine topography

One of the principal results of our analysis is an estimate of the pre-glacial channel steepness, \( k_s \) (Figure 4.10b), which should reflect climate, rock erodibility and rock uplift (Whipple and Tucker, 1999). We find that the highest pre-glacial \( k_s \) values are located in the western Alps, from the Mount Blanc massif to the Dent-Blanche and Sesia-Lanzo zones. The Pelvoux and Argentera crystalline massifs also show
4.4 Reconstruction of the pre-glacial k and Alpine topography

Figure 4.10: (a) Map of the estimated isostatic rock uplift in response to Quaternary erosion. The thick black line represents the base level. (b) Pre-glacial k map. The External Crystalline Massifs, the Dent Blanche and Sesia-Lanzo units and the Tauern and Engadine windows are shown in red, gray and blue, respectively. Black lines represent major tectonic lineaments. (c) Map of the reconstructed pre-glacial topography of the Alps. Elevations are above present-day sea level. The thick black line represents the base level. Note that, outside the base level, the present-day topography is displayed. (d) Map of the elevation difference, \( \delta \), between the present-day and modeled pre-glacial Alpine topographies. Green, red, pink, yellow, azure and brown basin outlines show the Rhone, Rhine, Aosta, Chiavenna, Adige and Puster Valleys, respectively. The white lines represent the LGM ice extent as compiled by (Buoncristiani and Campy, 2004).
4. Pre-glacial topography of the European Alps

high pre-glacial $k_s$ values. The Aar crystalline massif only shows high values in its western half. Other relatively high pre-glacial $k_s$ values are found in the Belledonne crystalline massif, the western half of the Tauern window, the Austro-alpine basement nappes and along the central part of the Insbruc line. Middle to low pre-glacial $k_s$ values are mostly found within Mesozoic and Cenozoic oceanic and continental sedimentary units. Finally, very low $k_s$ values are predicted toward the margins of the Alpine range and in a region to the south bounded by the Giudicarie, southern Engadine and Puster tectonic lines.

The reconstructed morphology of the Alps prior to Quaternary glaciation is shown in Figure 4.10c. The difference between the present-day and reconstructed topography, $\Delta z$ Figure 4.10d, shows negative values where erosion has exceeded rock uplift, indicating lowering of the surface topography over the period of glacial erosion. These areas are almost exclusively located within major Alpine valley bottoms. On the other hand, positive $\Delta z$ values, mostly located around topographic highs, depict regions where rock uplift has exceeded exhumation and indicate uplift of surface topography.

4.5 Discussion and Conclusions

The main feature of our inferred pre-glacial $k_s$ map is a region of high values in the western Alps which cuts across lithologic and tectonic province boundaries. The lack of evidence for young tectonic activity or lower precipitation rates restricted to this area leads us to infer a potential change in regional isostatic support of the western Alps. For example, it has been suggested that this region experienced recent uplift by detachment of the European slab (e.g., Lippitsch et al., 2003). This region also exhibits younger thermochonologic ages, suggesting higher exhumation rates over the last few Ma (e.g., Vernon et al., 2008).

The inferred pattern of pre-glacial $k_s$ across the rest of the range can also be related to rock uplift, climate or erosion patterns prior to glaciation. We find that high values of pre-glacial $k_s$ are often associated with harder rocks of the basement and metamorphic nappes, while low values often correspond to weaker Alpine flysch deposits (see also Figure 4.11). Given that there is little evidence for tectonic uplift of any isolated massifs in the Alps since 5 Ma (e.g., Willet,
4.5 Discussion and Conclusions

Insubric Line
Basal Alpine Thrust
External Crystalline Massifs
Penninic nappes
Austroalpine nappes
Southern Alps
Genova
Milano
Venezia
Grenoble
Zurich
Bern

Figure 4.11: Simplified tectonic map of the Alps from (Schmid et al., 2004).

2010, and references therein) we expect that this is primarily controlled by rock erodibility.

Figure 4.12: Topographic profiles (vertical exaggeration: 1:10) of the reconstructed pre-glacial (gray line) and present-day (full line) topographies. Elevations are above present-day sea level. The profile locations are shown in Figure 4.10d.

The reconstructed topography of the Alps shows narrower valleys and significantly lower maximum elevations prior to glaciation (Figure 4.10c). Our model
4. Pre-glacial topography of the European Alps

suggests that most glacial erosion has occurred in the lower and peripheral parts of the Alpine catchments (Figure 4.10d), below the long term glacial ELA. A ~1000 m surface lowering is obtained, for example, in the Rhone, Rhine, Adige and Puster valley bottoms and matches well with previous investigations (Jaboyedoff and Der-ron, 2005; Preusser et al., 2010). According to our preferred model, glacial erosion decreased mean elevation of the Alps from 1375±641 m to 1310±720 m (confirmed by a t-test at the 95% level) and increased valley scale topographic relief across the entire range, up to a factor of two (Figures 4.10d and 4.12), which is consistent with previous analyses (Haefelmann et al., 2007; Valla et al., 2011, 2012). This is partially a consequence of the base assumptions of fixed channel head elevation and concavity prior to glaciation. We discuss the sensitivity of our results to these assumptions below, but note that the selection of concavity is within the range of theoretical values of bedrock channels (i.e., 0.350.6 (Whipple and Tucker, 1999)) and modern concavity in the glaciated Alps is much higher than this range (Norton et al., 2010).

![Figure 4.13: Longitudinal profiles of the reconstructed pre-glacial (dotted line) and present-day (black line) Aosta, Rhone and Chiavenna valleys (vertical exaggerations: 1:20 for the Aosta valley and 1:25 for the Rhone and Chiavenna valleys). Elevations are above present-day sea level. The valley locations are shown in Figure 4.10d.](image)

This analysis shows that it is possible to resolve pre-glacial $k_s$ and elevations from modern channel head heights, however with a number of assumptions that should be assessed. Our analysis assumes that the Alpine river network was largely in a topographic steady-state at the onset of glaciation. By late Pliocene, the Alps had been elevated and eroding for nearly 30 Myr, and erosion rates were near-constant until at least the late Miocene, and possibly longer (Bernet et al., 2009). Although there is evidence for higher erosional fluxes from the Alps in the last 5 Myr (Kuhlemann et al., 2002), the temporal resolution of this signal is
not well resolved and much of this sediment flux could be the outcome of glacial erosion that we investigate here or recycling of older deposits (Sadler, 1999). Not surprisingly, sensitivity tests (Figures 4.7 and 4.8) show that the degree of valley deepening depends on the imposed channel concavity and lowering of channel head elevations during glaciation. However, the reconstructed pattern of pre-glacial $k_s$ (Figure 4.10b) is robust and independent of our assumptions. In addition, existing evidence from recent thermochronological studies suggests low erosion rates at high altitudes and increased valley incision (Glotzbach et al., 2011; Valla et al., 2011, 2012), in turn supporting our estimates of distribution and magnitudes of glacial erosion (Figures 4.10d, 4.12 and 4.13), increase of valley-scale topographic relief and decrease of mean elevation. Finally, it is possible that the effects of enhanced Quaternary erosion, associated isostatic rock uplift or sedimentation within major valleys result in a change of base level elevation. However, other processes often balance these effects. For example, deep carving of valleys below sea level is countered by sedimentation in the valley (Herman et al., 2011a); uplift of the surrounding foreland basins is countered by incision of rivers into the basin fill. The net effect should thus be within the uncertainty level of our analysis (i.e., $\sim$200 m, Figure 4.6).

In conclusion, we propose an advanced model of the pre-existing fluvial network elevation and use this to infer the late stage of rock uplift and erosion of the European Alps. The method presented here can be applied to reconstruct the pre-glacial topography of any alpine mountain range and will provide information on pre-glacial tectonic and climatic conditioning of mountain ranges.
Chapter 5

Geomorphic control and headward propagation of glacial erosion in the Rhône valley


5.1 Introduction

Quaternary glaciations have played a fundamental role in the evolution of high-elevation and mid- to high-latitude mountain ranges on Earth (e.g., Penck, 1905; Sugden and John, 1976). Glacial processes not only affect the morphology of the landscape (e.g., Harbor et al., 1988; Montgomery, 2002; Brocklehurst and Whipple, 2002), but also leads to complex feedbacks between tectonics, climate and erosion (e.g., Molnar and England, 1990; Raymond, 2004). Late Cenozoic global cooling (Zachos et al., 2001) seems to have increased the sediment yield (e.g., Zhang et al., 2001; Hay et al., 2002) and produced additional components of rock uplift (e.g., Molnar and England, 1990; Mitrovica and Peltier, 1991). Although some have questioned these effects (e.g., Willenbring and von Blanckenburg, 2010), it has been shown that glacial processes influence sediment fluxes (e.g., Piper et al., 1994;
Bowerman and Clark, 2011), recent exhumation histories (e.g., Shuster et al., 2005) and modern patterns of rock uplift (e.g., Pelletier, 2004; Spotila et al., 2004). However, the inferred increase in global sediment yield dates back to the late Pliocene, and it is too early to be attributed to Pleistocene glaciation only (Raymo, 1994). In addition, the overall effect of glacial erosion and associated isostatic adjustments on the topographic evolution of mountain ranges is not clear. For example, while large scale topographic analyses suggest that glaciation limits mountain growth by focusing erosion around the glaciers’ long-term Equilibrium Line Altitude (ELA) (e.g., Broecker and Denton, 1989a; Brozovic et al., 1997; Brocklehurst and Whipple, 2002; Egholm et al., 2009; Pedersen et al., 2010; Champagnac et al., 2011), measurements of long-term denudation rates and numerical models support increased erosion of lower reaches of the landscape with a consequent increase of topographic relief (e.g., Shuster et al., 2005; Herman et al., 2011b).

These apparent discrepancies are also observed in the European Alps which, given the historical and geological record of past glaciations, are often used as a natural laboratory to investigate how glacial processes affect an orogenic belt (e.g., Champagnac et al., 2007; Molnar, 2004; Norton et al., 2010). Measurements of sediment yields from the Alps are consistent with global observations (Hay et al., 1992; Kuhlemann, 2000; Kuhlemann et al., 2001), but the onset of glacial erosion is still debated. While in the Swiss and Bavarian foreland of the Alps glacial deposits date back to ∼2 Ma (Penck and Brückner, 1909; Schlichter and Kelly, 2000; Häuselmann et al., 2007), the earliest glacial deposits in the Po plain suggest a rather late onset of major glaciation at ∼0.9 Ma (Muttoni et al., 2003). Although the glacial imprint of the present-day Alpine topography in correspondence to the mean ELA is strong (Sternai et al., 2011) and pronounced glacial cirques suggest significant glacial erosion at even higher elevations (Anders et al., 2010), recent studies suggest a rapid relief development since the mid-Pleistocene (Haeuselmann et al., 2007; Glotzbach et al., 2011; Valla et al., 2011, 2012). This increase of topographic relief is also supported by numerical models of glacial erosion that include subglacial hydrology (Herman et al., 2011b) and paleo-topographic reconstructions (Szekely, 2003; Champagnac et al., 2007; Sternai et al., 2012).

In recent years, numerical modeling has been widely used to investigate processes of glacial erosion (e.g., Harbor et al., 1988; Tomkin and Braun, 2002; Amund-
son and Iverson, 2006; Herman and Braun, 2008; Herman et al., 2011b; Egholm et al., 2011, 2012). A common result is that initial hypsometry (i.e., the distribution of elevations) has a significant effect on the patterns and magnitudes of glacial erosion (Braun et al., 1999; Egholm et al., 2009; Sternai et al., 2011). However, the morphology of a landscape before the onset of glaciation is rarely known. Recently, Sternai et al. (2012) proposed a new method to reconstruct the pre-glacial topography of the European Alps. This study provides a unique opportunity to evaluate the efficiency of glacial erosion on such a pre-glacial landscape. By comparing the patterns and magnitudes of glacial erosion on such a topography to those obtained on the present-day landscape, we can estimate the spatial and temporal evolution of glacial erosion in the Alps.

This study is focused on the Rhône valley, Switzerland, one of the major drainage systems of the Alps. Three numerical experiments were specifically designed to investigate the role of landscape morphology in setting ice building and glacial erosion patterns, the effects of several glacial cycles on the morphologic evolution of an initially fluvial landscape and the extent to which the period of climate oscillations affects the sediment yield. In the following, we first describe the geologic, geomorphic and geodetic observations that we used to constrain these numerical experiments. We then outline the modeling strategy, provide an overview of the numerical model and depict the model calibration. Finally, we present and discuss the model results. Major outcomes suggest that glacial erosion is more complex than a simple “buzz-saw” effect (Brozovic et al., 1997; Mitchell and Montgomery, 2006; Egholm et al., 2009) or increase of topographic relief (e.g., Haeselmann et al., 2007; Glotzbach et al., 2011; Valla et al., 2011, 2012). We find that glacial erosion propagates headward through time, leading to an initial increase of topographic relief which is followed by erosion at high elevations.

5.2 Setting - the Rhône valley

In this section, we outline the major geologic, geomorphic and geodetic observations that have directed our study. We only emphasize aspects that are most appropriate for this analysis, while further details can be found in (e.g., Florineth and Schlüchter, 1998; Kelly et al., 2004; Jaboyedoff and Derron, 2005; Schlatter
et al., 2005; Ivy-Ochs et al., 2006a; Norton et al., 2010).

5.2.1 Geomorphology and glacial record

The Rhône valley (Figure 5.1) is one of the major drainage systems of the Alps. While the principal architectural elements of the Rhône valley were already established by the end of the Oligocene (e.g., Schmid et al., 2004), Pleistocene glaciation and associated isostatic adjustment have heavily impacted this valley. Glacial erosion has generated a high relief landscape with peaks higher than 4000 m and several hundred meters deep over-deepenings (Jaboyedoff and Derron, 2005; Preusser et al., 2010). Glacial erosional features such as trimlines, striae or polished rocks are abundant above the ELA and the valley bottom mostly preserves the classical “U”-shape. Erratic boulders also occur within the Rhône valley and provide information about ice extent and flow directions (Kelly et al., 2004; Ivy-Ochs et al., 2006a). Relatively little information about most of the fourfold glaciation system of the Alps (e.g., Buoncristiani and Campy, 2004) is preserved within the study region. Prominent moraine systems and associated deposits formed in the valleys and cirques during the Last Glacial Maximum (LGM), however, provide an important morphological reference along the Rhône valley. Reconstructions of LGM ice extent show that the Rhône valley was covered by up to ~1500 m of ice (Florineth and Schlüchter, 1998; Kelly et al., 2004). By ~21 kyr before present the Rhône valley glacier abandoned the terminal moraines (Ivy-Ochs et al., 2006a). Although such glacial reconstructions and their relations with past climate are relatively well established, the physical processes controlling ice building and glacial erosion are still elusive.

Recently, Sternai et al. (2012) have presented a reconstruction of the pre-glacial Alpine landscape which accounts for spatially variable channel steepness and estimates of the isostatic adjustment to Quaternary erosion. By comparing such a pre-glacial landscape to the present-day topography, they constrain patterns and magnitude of glacial erosion and observed an increase of local relief throughout glaciation, in spite of a decrease of mean elevation. This is also verified in the Rhône valley, where the hypsometries before and after glaciation show the development of high peaks, lowering of the valley bottom and the formation of an
over-deepened trough as a result of glacial erosion (Figure 5.2).

5.2.2 Rock uplift

Rock uplift can enhance the accumulation area of glaciers in turn affecting ice building, glacial erosion and sediment yield. Variations of rock uplift and exhumation rates have occurred during the late Neogene and Quaternary, for example across the Penninic line (Vernon et al., 2008; Sternai et al., 2012). However, the magnitude and timing of these changes are not sufficiently well resolved to provide information about the evolution of rock uplift patterns throughout glaciation. GPS data indicate that less than 2 mm yr\(^{-1}\) of integrated convergence rate with respect to stable Europe is currently accommodated in the western Alps and strain is es-
5.2 Setting - the Rhône valley

Figure 5.2: Red bars represent the hypsometry of the Rhône valley prior to glaciation (as reconstructed by Sternai et al. (2012)). Black bars show the hypsometry of the Rhône valley at the end of the second numerical experiment with null tectonic rock uplift (see text for details). Dotted black bars represent the present-day hypsometry of the Rhône valley corrected for the sediment cover (based on estimates of post glacial sediment infill by Jaboyedoff and Derron (2005)). The gray envelop represents LGM and modern ELAs as estimated by Ivy-Ochs et al. (2006a) or compiled by the World Glaciers Monitoring Service (IUGG/UNEP/UNESCO, 2005).

essentially controlled by the counter-clockwise rotation of the Adriatic microplate with respect to Eurasia (Calais et al., 2002; D’agostino et al., 2008). Vertical displacements relative to a reference point in Aarburg (Swiss Molasse basin) were measured as $\sim 0.25 - 1.5$ mm yr$^{-1}$ (Schlatter et al., 2005) and are probably the response to the ongoing collision between the Adriatic plate system and the European lithosphere (Kahle et al., 1997; Calais et al., 2002; D’agostino et al., 2008) and the isostatic re-adjustment to erosion (e.g., Champagnac et al., 2007) and recent ice melting (Barletta et al., 2006). The relative importance of these three components in determining measured rock uplift rates, however, is still a matter of discussion.
5.3 Numerical analysis

In this section, we first describe our modeling strategy, then provide a brief overview of the numerical model and, finally, report the model calibration.

5.3.1 Modeling strategy

Numerical experiments were designed to address (A) the influence that landscape morphology has on ice building and glacial erosion patterns, (B) the effects of several glacial cycles on the morphologic evolution of an initially fluvial landscape and (C) the extent to which the period of climate oscillations affects the sediment yield. As detailed below, we used observations regarding various aspects of glacia tion in the Rhône valley to calibrate our numerical model and produce realistic simulations of glacial erosion in this valley. We remark, however, that it is not our purpose to reproduce the full evolution of the study region from pre-glacial times to modern days, nor to thoroughly explain its present-day geomorphology, but rather to assess major modifications of the topography by early and late glacial erosion in the Rhône valley. For this reason, fluvial and hillslope processes in the landscape evolution model were minimized.

5.3.2 Numerical model

We used a modified version of the ICE-CASCADE code. We provide here a brief overview of the governing equations, while more details about the algorithm can be found in (Braun et al., 1999; Tomkin and Braun, 2002; Herman and Braun, 2008; Herman et al., 2011b).

Ice model

The ice thickness, $h$, is calculated by solving the equation of ice mass conservation:

$$\frac{\partial h}{\partial t} = \nabla \cdot Q + M$$  \hspace{1cm} (5.1)

where $t$ is the time, $Q$ is the ice flux and $M$ is the ice net mass balance. $M$ is calculated as a linear function of the temperature, $T_s$, which scales linearly with
5.3 Numerical analysis

elevation:

\[ M = -\gamma T_s \quad \text{with} \quad T_s = T_0 - \lambda (h + z) \quad (5.2) \]

where \( \gamma \) and \( \lambda \) are arbitrary constants, \( z \) is the elevation and \( T_0 \) is the temperature at sea level that varies through time to simulate glacial cycles. Minimum \( (M_{\text{min}}) \) and maximum \( (M_{\text{max}}) \) rates of ice accumulation and ablation can also be set to reproduce a realistic ice net mass balance. To describe ice sliding velocities, \( u_s \), we apply a commonly used empirical law (Bindschadler, 1983; van der Veen, 1987; Paterson, 1994):

\[ u_s = \frac{B_s \tau^{m}}{p_e} \quad (5.3) \]

where \( \tau_b \) is the ice basal shear stress, \( B_s \) is a sliding constant, \( m \) is an arbitrary exponent and \( p_e \) is the effective pressure (i.e., \( p_i - p_w \), where \( p_i \) is the ice overburden pressure and \( p_w \) is the water pressure).

**Subglacial hydrology model**

Following Flowers and Clark (2002), the water thickness, \( h_w \), is computed as:

\[ \frac{\partial h_w}{\partial t} = b + \dot{b}_b - \nabla \cdot Q_w \quad (5.4) \]

where \( Q_w \) is the water discharge, \( \dot{b} = |\min(M, 0)| \) is the meltwater produced by ablation and \( \dot{b}_b \) is a background basal melt. The computation of the water pressure is based on an empirical formula:

\[ p_w = p_i \left( \frac{h_w}{h_c} \right)^\alpha \quad (5.5) \]

where \( h_c \) is the critical thickness of the confined aquifer and \( \alpha \) is an empirical exponent equal to 3.5 (Flowers and Clark, 2002).

**Glacial erosion model**

Landscape evolution models often assume that glacial erosion is dominated by abrasion and quarrying (e.g., Hallet, 1979, 1996; MacGregor et al., 2000; Anderson et al., 2006; Herman and Braun, 2008; Egholm et al., 2009; Pelletier et al., 2010)
5. Geomorphic control and headward propagation of glacial erosion in the Rhône valley

Table 5.1: Parameters settings. The values are applied to all numerical experiments, unless otherwise specified in the text.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$B_s$</td>
<td>$5 \times 10^{-13}$ Pa$^{-3}$ m$^{-1}$</td>
<td>sliding law parameter</td>
</tr>
<tr>
<td>$\rho$</td>
<td>910 kg m$^{-3}$</td>
<td>density of ice</td>
</tr>
<tr>
<td>$g$</td>
<td>9.81 m s$^{-2}$</td>
<td>gravitational acceleration</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>2.</td>
<td>mass balance slope constant</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>0.006°C m$^{-1}$</td>
<td>lapse rate</td>
</tr>
<tr>
<td>$T_0$ min, $T_0$ max</td>
<td>9, 18°C</td>
<td>Min, Max sea surface temp.</td>
</tr>
<tr>
<td>$m$</td>
<td>3.</td>
<td>exponent sliding law</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>3.5</td>
<td>exponent water pressure</td>
</tr>
<tr>
<td>$l$</td>
<td>1.</td>
<td>exponent erosion law</td>
</tr>
<tr>
<td>$r_u$</td>
<td>0. - 0.5 mm yr$^{-1}$</td>
<td>tectonic rock uplift</td>
</tr>
<tr>
<td>$k_g$</td>
<td>2.e-5 - 1.e-5</td>
<td>glacial erosion coefficient</td>
</tr>
<tr>
<td>$\nu$</td>
<td>0.2</td>
<td>Poisson’s ratio</td>
</tr>
<tr>
<td>$T_e$</td>
<td>20 km</td>
<td>crustal elastic thickness</td>
</tr>
<tr>
<td>$t$</td>
<td>0.1 - 0.9 - 2 Myr</td>
<td>simulated time</td>
</tr>
<tr>
<td>$P$</td>
<td>40 - 100 kyr</td>
<td>period of climate oscillations</td>
</tr>
<tr>
<td>$t_s$</td>
<td>0.01 yr</td>
<td>ice time step</td>
</tr>
<tr>
<td>$t_e$</td>
<td>100 yr</td>
<td>erosion time step</td>
</tr>
</tbody>
</table>

and that these two processes are a function of the sliding velocity. A similar approximation is adopted here:

$$\frac{\partial z}{\partial t} = k_g u_s^l$$  (5.6)

where $k_g$ is a constant glacial erosion coefficient and $l$ is equal to 1. Erosion rates are scaled on the cosine of the bedrock slope to avoid runaway effects (e.g., Herman et al., 2011b).

5.3.3 Model calibration

We provide below a description of the data we used to calibrate our numerical model. A summary of the parameter setting is shown in Table 5.1.

Shape of climate oscillations and ice net mass balance

Oxygen isotope measurements on benthic foraminifera (Zachos et al., 2001) confirmed that modulations of climate coincide with Milankovitch cycles, as also pro-
posed by (Fairbridge, 1961; Broecker, 1966). Recurrent periods of oscillation are \( \sim 100 \) kyr and 40 kyr. While 40 kyr oscillations are mostly symmetric, those with a period of 100 kyr exhibit a sawtooth shape (e.g., Broecker and van Donk, 1970; Fairbridge, 1972). We thus approximate the 40 kyr cyclicity as a sinusoidal oscillation of the temperature at sea level and impose, for the 100 kyr oscillations, cooling and warming trends covering respectively 80% and 20% of each cycle (Figure 5.3). Maximum and minimum sea surface temperatures \( (T_0) \) were set to 18°C and 9°C, to match modern high-resolution observations of sea surface temperatures around Europe (Reynolds and Smith, 1995) as well as modern and LGM ELA estimates (i.e., 2900±150 m (1\( \sigma \)) and 1700±200 m (1\( \sigma \)), respectively, as detailed in IUGG/UNEP/UNESCO (2005); Ivy-Ochs et al. (2006a)). Assuming that Alpine glaciers were fed predominantly by snowfall, with a clean ablation area, we set \( M_{\text{max}} \) and \( M_{\text{min}} \) equal to 1 m yr\(^{-1} \) and -20 m yr\(^{-1} \) (Benn and Lehmkuhl, 2000), respectively. It is worth pointing that we found a rather good agreement between model predictions and independent reconstructions of LGM ice extent by Kelly et al. (2004) and Schlüchter et al. (2009) (Figure 5.4).
Erosional efficiency

The over-deepened troughs that surround the Alps are probably the most prominent erosional features of the belt (Preusser et al., 2010). Although the processes responsible for the formation of such noticeable depressions are still debated (e.g., Oerlemans, 1984; MacGregor et al., 2000; Kessler et al., 2008; Herman et al., 2011b), it is widely accepted that Alpine over-deepenings have glacial origins (e.g., Penck, 1905; Jaboyedoff and Derron, 2005; Preusser et al., 2010). The maximum depth of the over-deepened trough within the Rhône valley has been estimated by Jaboyedoff and Derron (2005) as $\approx 1000$ m below the present-day topographic surface. This estimate provide us with the possibility to calibrate the erosional efficiency of the modeled glacier. Referring to grid nodes away from the boarders to avoid edge effects on erosion estimates, we tuned the glacial erosion coefficient,
5.4 Results

$k_g$, to obtain erosion rates that produce a trough of similar depths over 1 to 2 Myr temporal scales.

Tectonic and isostatic rock uplift

As previously discussed, measurements of late Quaternary rock uplift may combine different components associated with isostatic adjustment to Quaternary erosion and recent ice melting, as well as a tectonic component of rock uplift due to ongoing orogenic processes (Champagnac et al., 2009). We assessed the isostatic components of rock uplift using a two-dimensional flexural model that computes the vertical deflection of an elastic plate produced by removal of the eroded material and changes in ice load through time. We used spatially uniform crust and mantle densities and set Poisson’s ratio and the elastic thickness of the crust according to previous estimates (Christensen, 1996; Stewart and Watts, 1997). To adjust modeled rock uplift to measured magnitudes (i.e., $\sim 0.25 - 1.5$ mm yr$^{-1}$ (Schlatter et al., 2005)), we run experiments for tectonic rock uplift equal to 0 and 0.5 mm yr$^{-1}$, constant in both space and time.

5.4 Results

We present the model results in this section. We performed three numerical experiments, each consisting of two simulations (as summarized in Table 5.2). In the numerical experiment A, we compare glacial erosion over one asymmetric glacial cycle (100 kyr) on the pre-glacial and present-day topography (simulations 1 and 2, respectively). The aim of this numerical experiment is to define the role of the initial morphology in setting patterns and magnitudes of glacial erosion. In the numerical experiment B, we explore how the landscape shifts from fluvial to glacial. We simulate nine glacial cycles on the pre-glacial landscape and examine the evolution of glacial reshaping and consequent changes in glacial erosion patterns. In order to identify the role that tectonic rock uplift has on such an evolution of glacial reshaping and erosion patterns, the simulations included in this numerical experiment (simulations 3 and 4) account for tectonic rock uplift rates equal to 0 and 0.5 mm yr$^{-1}$, respectively. In the numerical experiment C, we
investigate to which extent a shift of the period of climate oscillations from 40 to 100 kyr affect the ice volume, the average sliding velocity and the glacial sediment yield. Similarly to the second numerical experiment, we modulate tectonic rock uplift rates (equal to 0 and 0.5 mm yr$^{-1}$ in simulation 5 and 6, respectively) so that it was possible to define its effects on the inferred quantities.

In order to describe the magnitudes and patterns of glacial erosion, we calculate the glacial sediment yield through time and produce erosion maps of specific time steps. In addition, we selected 11 control points along the major valley trunk and higher lateral valleys and track the temporal evolution of glacial erosion in this key locations.

5.4.1 Numerical Experiment A (simulations 1 and 2): the role of the initial morphology

In this numerical experiment, we simulate one glacial cycle (100 kyr) on both the pre-glacial (simulation 1) and present-day (simulation 2) topography. The channel concavity of glacial catchments from the present-day topography are statistically indistinguishable and its average is $\sim$0.73 (Norton et al., 2010). The pre-glacial landscape was modeled imposing a uniform channel concavity of 0.5 (Sternai et al., 2012) to match standard concavity indices of fluvial landscapes (Whipple and Tucker, 1999). Thus, the modeled pre-glacial topography has a different hypsometry compared to the present-day landscape, with a higher frequency of elevations around the mean long term ELA (Figure 5.2). This results in a net increase of accumulation area on the pre-glacial landscape compared to the present-day topography. In turn, the volume on the pre-glacial topography is higher than on the present-day landscape by approximately a factor of 2 during
5.4 Results

the cooling and warming phases of the glacial cycle (Figure 5.5).

During initial cooling, glaciers on the present-day topography are restricted to the upper parts of the catchments where they concentrate erosion (Figures 5.6). In contrast, the ice cap on the pre-glacial topography develops rapidly to fill and erode the major valley trunk. This is further illustrated by the evolution of glacial erosion at the control points (Figure 5.7). Dashed lines show that glacial erosion is more effective on the upper catchments of the present-day topography, indicating that magnitudes and patterns of glacial erosion depends on the hypsometry of the landscape. Furthermore, the present-day landscape is characterized by particularly high slopes above the snow-line (Pedersen et al., 2010) which provides an additional contribution to increase ice sliding velocities and thus enhance glacial erosion in the high reaches of the landscape.

During the glacial maximum, water due to melting abounds in the ablation
5. Geomorphic control and headward propagation of glacial erosion in the Rhône valley

Figure 5.6: Selected snapshots of the simulation 1 and 2 (see Table 5.2). The location of the control points, color coded by elevation, is also shown in (a) and (b). See texts for further details.
area, resulting in a substantial increase of the sliding velocities (and thus erosion rates) at lower elevations (Figure 5.6e-f). Glacial erosion of the lower parts of the landscape is approximately one order of magnitude higher than that at high elevations on both topographies (Figure 5.7).

Similarly to the early stages of the glacial cycle, when glaciers retreat, the upper reaches of the landscape are subject to a higher glacial erosion on the present-day topography than on the pre-glacial landscape. Over a single glacial cycle, glacial reshaping of the pre-glacial topography occurs almost exclusively towards lower reaches of the landscape (Figure 5.6g), while a bimodal distribution of glacial erosion with a peak at high elevation and enhanced erosion at low altitudes is found on the present-day landscape (Figure 5.6h).

Figure 5.7: Glacial erosion through time for each control point (see Figure 5.6 for control point locations) on the (a) pre-glacial and (b) present-day topography. The color coding represents the elevation of the control points. Full and dashed lines on both panels represent glacial erosion through time for points in the main valley trunk and lateral valleys, respectively. Glacial erosion at low elevation (light and dark blue lines on both panels) is approximately one order of magnitude higher and follows the scale on the right. Glacial erosion of the main valley trunk (full lines on both panels) is inversely proportional to elevation. Glacial erosion of present-day lateral valleys (dashed lines in (b)) is significantly higher than what is found on equivalent valleys of the pre-glacial landscape (dashed lines in (a)).
5.4.2 Numerical Experiment B (simulations 3 and 4): long-term evolution of glacial erosion

In this numerical experiment, we address glacial erosion since the Mid-Pleistocene transition (Muttoni et al., 2003). We thus simulate nine glacial cycles (100 kyr) on the pre-glacial landscape, using a component of tectonic rock uplift equal to 0 and 0.5 mm yr\(^{-1}\) (simulations 3 and 4, respectively). The evolution of glacial erosion over several glacial cycles is characterized by a headward propagation of glacial erosion through time, independent of the selected tectonic rock uplift rate (Figure 5.8 and 5.9). This headward propagation of glacial erosion is also shown by the total amount of erosion and erosion rates of each control point (Figure 5.10). The initial valley carving at low elevations steepens the slopes in the upstream part of the trough with a consequent increase of the ice sliding velocities during the subsequent glacial cycle. This mechanism has the double effect of decreasing erosion rates at low elevations and triggering headward propagation of erosion during several glacial cycles. After 0.9 Myr of glacial erosion, an over-deepened trough with similar location and depth to the one observed on the present-day landscape is formed.

Climatic oscillations control the ice extent, the velocity pattern and, thus, erosion rates (Figure 5.11). A close examination of erosion during a single glacial cycle reveals that sediment production rates are higher during the phases of ice advance and retreat, when the glacier is rapidly changing its shape and extent. There is also a decrease of the sediment yield through time, when the simulation does not include any component of tectonic rock uplift. Such a decrease is expected as it is the consequence of a reduction of glacier’s accumulation area. In contrast, we find that a tectonic rock uplift of 0.5 mm yr\(^{-1}\) compensates the loss of accumulation area caused by glacial erosion. The ice volume at the end of each glacial cycle, in fact, increases through time (Figure 5.9) and a constant sediment yield is maintained throughout the entire simulation (Figure 5.11b).

Our numerical study includes a number of simplifications that preclude to fully reproduce the hypsometric differences that are observed between the pre-glacial and present-day topography (Figure 5.2). For example, modeled erosion around the long term ELA is underestimated. This might be due to the fact
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Figure 5.8: Selected snapshots from simulation 3 (see Table 5.2). Glacial erosion patterns after each glacial cycle on the pre-glacial topography with no tectonic components of rock uplift. The blue line in each panel represents an isoline of erosion equal to 100 m and highlights headward propagation of the erosional front through the glacial cycle. The location of the control points, color coded by elevation, is also shown in the top-left panel.
5. Geomorphic control and headward propagation of glacial erosion in the Rhône valley

Figure 5.9: Selected snapshots from simulation 4 (see Table 5.2). Glacial erosion patterns after each glacial cycle on the pre-glacial topography with tectonic rock uplift equal to 0.5 mm yr$^{-1}$. Other features are the same as Figure 5.8. Note the headward propagation of glacial erosion and the increasing ice extent at the end of each glacial cycle induced by tectonic uplift.
5.4 Results

Figure 5.10: Glacial erosion (a) and erosion rates (b) after each of nine glacial cycles of simulations 3 and 4 (see Table 5.2), for control points in the main valley trunk (see the top-left panel in Figure 5.8 or 5.9 for control point locations). Glacial erosion at low elevation (light and dark blue dots on both panels) follows the scale on the right. Full and dashed lines represent the evolution of glacial erosion for points in the main valley trunk and lateral valleys, respectively.

that we do not include neither fluvial nor periglacial erosion during interglacial periods. Other contributions that we do not account for are the effects of loose sediments on bedrock erosion (e.g., Boulton, 1979, 1996), sediment transport and variations of rock erodibility across the study area. Nonetheless, the hypsometry of the landscape at the end of this numerical experiment approaches the one of the present-day landscape (Figure 5.2), in turn indicating that the numerical model is able to reproduce the major effects of glacial erosion.

5.4.3 Numerical Experiment C (simulations 5 and 6): shift of the period of climate oscillations

In this numerical experiment, we simulate 2 Myr of glacial erosion (Penck and Brückner, 1909; Schlüchter and Kelly, 2000; Häuselmann et al., 2007) on the pre-glacial topography, imposing a shift of the period of climate oscillations from 40 kyr to 100 kyr at 1 Myr (Figure 5.12, Zachos et al., 2001; Lisiecki and Raymo, 2007). We compare the evolution of the ice volume, average ice sliding velocity and glacial sediment yield through time, for different simulations with tectonic rock uplift
5. Geomorphic control and headward propagation of glacial erosion in the Rhône valley

Figure 5.11: (a-b) Black lines show the temporal evolution of sediment yield for simulations 3 and 4 (see Table 5.2). Sediment yield is maximized during the advancing and retreating phases of the glacier, suggesting that the sediment yield is independent on the ice volume or extent. Red dots and lines represent the results averaged over each glacial cycle.

Figure 5.12: Modeled climate oscillations for simulations 5 and 6 (see Table 5.2). During the first Myr, glacial cycles are approximated by sinusoidal oscillations with period equal to 40 kyr. After 1 Myr, the period shifts to an asymmetric cycle (warming and cooling phases are 20% and 80%, respectively) with period equal to 100 kyr.

equal to 0 and 0.5 mm yr\(^{-1}\) (simulations 5 and 6, Figure 5.13). Similarly to the second numerical experiment, a tectonic rock uplift rate of 0.5 mm yr\(^{-1}\) is sufficient to counterbalance the loss of accumulation area induced by glacial erosion and increase the ice volume through time. This increase of ice volume is accompanied by an increase of the average ice sliding velocity and, thus, of glacial erosion and

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sediment yield. In contrast, when the tectonic component of rock uplift is set
to 0 mm yr\(^{-1}\), the amount of ice decreases through time and a similar trend is
found for all other parameters. The dependency that links ice volume, sliding
velocity and sediment yield to tectonic rock uplift is even more evident when these
quantities are averaged over single glacial cycles (red line in Figure 5.13). The
variability of these quantities over single glacial cycles is roughly constant when
the tectonic component of rock uplift is not included. However, model boundary
conditions prevent any consideration about the variability of ice building through
time when tectonic rock uplift is equal to 0.5 mm yr\(^{-1}\) as the ice coverage is limited
by the grid size.

The major effect of shifting from short to long climate oscillations is an increase
of ice volumes, sliding velocities and sediment yield (Figure 5.13). According to
our results:

\[ \int_{t_0}^{t_1} V dt < \int_{t_1}^{t_2} V dt \]  \hspace{1cm} (5.7)

where \(V\) is the ice volume within the study region, \(t_0 = 0.9\) Myr, \(t_1 = 1\) Myr
(the time at which the shift of the period of climate oscillations occurs) and \(t_2 =
1.1\) Myr. In particular, changing the period of climatic oscillations from 40 kyr to
100 kyr increases the integrated ice volume by \(\sim 15\%\) on both simulations 5 and
6. A similar increase is also found for the average sliding velocity and sediment
yield. A shift from short to long climate oscillations thus extends the time spans
during which glacial erosion occurs and promotes the production of sediments
due to glacial erosion. After few long period glacial cycles, however, this effect is
dampened suggesting a rather fast re-equilibration of the erosive regimes to the
new conditions and variations of ice volume, average sliding velocity and sediment
yield are again driven by tectonic rock uplift (as described above).

## 5.5 Discussion

Global analyses of topography show that variations in mountain height corre-
lates with present-day and LGM snow-lines (Broecker and Denton, 1989b; Egholm
et al., 2009) and are independent of tectonic convergence rates (Champagnac et al.,
2011). In fact, the hypsometry of modern glaciated mountain ranges suggests that
denudation mechanisms are more effective at high elevations, leading to the conclusion that glaciation limits topographic relief and mountain growth (e.g., Brozovic et al., 1997; Montgomery et al., 2001; Mitchell and Montgomery, 2006; Foster et al., 2008; Pedersen et al., 2010; Sternai et al., 2011). The present-day landscape, however, is more prone to bear information about the most recent stages of glacial erosion, as erosion itself erases evidence of past processes and landforms. Different patterns of glacial erosion on the pre-glacial and present-day topography suggest that glacial effects are not as simple as a “buzz-saw” erosion (e.g., Brozovic et al., 1997; Mitchell and Montgomery, 2006; Egholm et al., 2009) or increase of topographic relief (e.g., Haeuselmann et al., 2007; Glotzbach et al., 2011; Valla et al., 2011, 2012), but rather that they evolve through time, also in relation to the evolution of the landform. Our results show that glacial erosion is focused toward the lower reaches of the landscape during the early stages of glaciation (Figures 5.6g and 5.7a). Early glacial erosion, thus, produces an increase of topographic relief, consistent with modern glacial over-deepenings. By dating valley fill deposits within glacial over-deepened troughs in the proximity of the Rhône valley, Preusser et al. (2010) provided a minimum age for the formation of these features \( \sim 250 \text{ kyr} \). Similarly, Haeuselmann et al. (2007) and Valla et al. (2011) found that the major phase of glacial erosion in the Aar and Rhône river basins occurred at \( \sim 0.9 \text{ Myr} \) and increased topographic relief. These findings provide support to our results. During the late stages of glaciation, however, glacial erosion impacts the highest reaches of the landscape (Figures 5.6h and 5.7b), a result that matches well with geomorphic investigations of the present-day Alpine topography (Sternai et al., 2011).

As the ice sliding velocities and, thus, the rates of glacial erosion are proportional to local slopes, the erosive power of the glacier is particularly high when it flows down a steeper slope formed during the initial carving of the over-deepened trough. This effect triggers a headward propagation of glacial erosion (Figure 5.8), similar to knickpoints in fluvial domains that migrate from the base-level towards the headwaters. Headward propagation of glacial erosion was already proposed to explain thermochronometric data from the mountainous landscapes of Fiordland, New Zealand (e.g., Shuster et al., 2005). Even more importantly, headward propagation of glacial erosion is the only class of modeling results capable of matching
evidence for a glacially induced increase of topographic relief (Valla et al., 2011) within the Rhône valley as well as increased glacial erosion in correspondence of the long-term ELA (Sternai et al., 2011) (see also Figure 5.2). Interestingly, our result suggests that glacial valley deepening and plane off at and above the long term-ELA are related processes, linked by headward propagation of glacial erosion through time.

The review of global field data by Hallet et al. (1996) supports the idea that fast-moving glaciers can be more erosive than rivers. An increase of sediment yield associated to a cooling climate is therefore to be expected. The temporal resolution provided by detailed studies of sediment yield from the Alps (for example those by Kuhlemann (2000) or Kuhni and Pfiffner (2001)), however, is not sufficient to evaluate how the sediment production by glacial erosion evolves throughout glaciation. Our results suggest that the sediment yield decreases through time if no tectonic rock uplift compensates for the loss of accumulation area induced by glacial erosion (Figure 5.11 and 5.13). However, a tectonic component of rock uplift of 0.5 mm yr$^{-1}$ is sufficient to increase the sediment yield throughout glaciation which is in agreement with measurements of sediment yields from the Alps (Hay et al., 1992; Kuhlemann, 2000; Kuhlemann et al., 2001) and thermochronometric data (Valla et al., 2011, 2012). It is worth noting, however, that an additional cooling trend superimposed to the glacial cycles or an increase of the amplitude of the latter might equally compensate for the loss of accumulation area induced by glacial erosion and, thus, increase the sediment yield. The lack of fluvial and hillslope processes in our numerical experiments might emphasize this effect. Yet this behavior indicates that, when climate conditions allow glaciations, tectonic rock uplift or long term climatic trends have a primary control on the production of sediments.

A shift of the period of climate oscillations from 40 kyr to 100 kyr extends the time spans during which glacial erosion occurs and promotes the production of sediments (Figure 5.13). Thus, effects such as lengthening of the period of climate oscillations might play a role in setting the efficacy of glacial erosion (Figure 5.13). This is important considering, for example, the discrepancy between the age of the earliest glacial deposits in the northern foreland (i.e., ∼2 Myr (Penck and Brückner, 1909; Schlüchter and Kelly, 2000; Häuselmann et al., 2007)) and
5. Geomorphic control and headward propagation of glacial erosion in the Rhône valley

in the Po plain (i.e., \( \sim 0.9 \) Myr (Muttoni et al., 2003)). Given that there is little evidence for any significant change of tectonic rock uplift rate since the late Pleistocene (e.g., Willett, 2010, and references therein), our numerical results suggest that the climatic transition from symmetric 40 kyr to asymmetric 100 kyr periods of climate oscillations during the mid-Pleistocene (Lisiecki, 2010) might have enhanced glacial erosion rates by extending the time spans of glacial erosion and promoting new transient conditions. Another possibility is that a change in regional isostatic support of the Western Alps (i.e., slab break-off (Lippitsch et al., 2003)) has increased rock uplift rates with consequent augmentation of glaciers’ accumulation area and thus glacial sediment yield. These processes also provides possible explanations for geochronometric datasets suggesting an increase of valley deepening at \( \sim 1 \) Myr in the Aar river basin (Haeuselmann et al., 2007) and the Rhône valley (Valla et al., 2011).

5.6 Conclusion

Simulations of glacial erosion on the pre-glacial and present-day landscape calibrated on observational constraints within the Rhône valley enabled us to produce results leading to the following conclusions:

- The effects of glaciation are more complex than a simple “buzz-saw” effect or increase of topographic relief. Glacial conditioning of a landscape evolves through time, also in relation with the evolution of the landscape itself. Patterns of glacial erosion on the pre-glacial and present-day topography of the Rhône valley suggest an initial increase of topographic relief followed by enhanced erosion at higher elevations during late stages of glaciation.

- Valley over-deepening and enhanced glacial erosion at and above the long term ELA of the Rhône valley are linked by a headward propagation of glacial erosion through time. Such a headward propagation of glacial erosion is capable of matching evidence for a glacially induced increase of topographic relief (Valla et al., 2011, 2012) within the Rhône valley as well as increased glacial erosion in correspondence of the long term ELA (Sternai et al., 2011).

- In absence of any tectonic component of rock uplift, additional cooling trends superimposed to glacial cycles or increase of the amplitude of climate oscillations
to compensate for the decrease of glaciers’ accumulation area induced by glacial erosion, glacial sediment yield decreases during successive glacial cycles. If climate oscillations are constant through time and no additional, long term cooling trends are included, a tectonic component of rock uplift of 0.5 mm yr$^{-1}$ is sufficient to increase the production of sediments throughout glaciation, in agreement with measurements of sediment yields from the Alps (Hay et al., 1992; Kuhlemann, 2000; Kuhlemann et al., 2001).

- A shift of the period of climate oscillations from 40 kyr to 100 kyr can enhance the sediment yield by extending the time spans of glacial erosion and promoting new transient conditions. This mechanism or a change in regional isostatic support of the Western Alps (Lippitsch et al., 2003) are likely factors to explain the discrepancy between the age of the earliest glacial deposits in the northern foreland (Penck and Brückner, 1909; Schlüchter and Kelly, 2000; Häuselmann et al., 2007) and in the Po plain (Muttoni et al., 2003) as well as thermochronometric datasets suggesting an increase of valley deepening by glacial erosion at ~1 Myr (Haeuselmann et al., 2007; Valla et al., 2011).
5. Geomorphic control and headward propagation of glacial erosion in the Rhône valley

Figure 5.13: Black lines represent the temporal evolution of the ice volume (a-b), sliding velocity (c-d) and sediment yield (e-f). Red lines represent the results averaged over each glacial cycle. Left-hand-side panels show results for the experiment with no tectonic component of rock uplift (simulation 5, see Table 5.2). Right-hand-side panels show results with 0.5 mm yr$^{-1}$ of tectonic rock uplift (simulation 6, see Table 5.2). After 1 Myr, the period of glacial cycles shifts from 40 kyr to 100 kyr (vertical dashed line).
Chapter 6

Conclusions

During the time of this thesis, I developed new techniques and applied well known methodologies to extend our knowledge of the recent topographic evolution of the European Alps. The main conclusions are summarized as follows:

- An examination of the effects that glacial erosion has on the hypsometric curve of idealized cross-sectional valley profiles highlights that very different valley geometries can have a similar hypsometric integral, \( HI \). The same analysis suggests that the gradients of the hypsometric curve may provide less ambiguous insights into the amount of glacial erosion experienced by a landscape. In order to resolve for the glacially induced modifications on hypsometry, I defined a new morphometric parameter, called the \( \text{hypsokyrtome} \), \( H_{kr} \), as the minimum normalized elevation at which the gradient of the hypsometric curve is greater or equal to a reference value \( (S_f) \). Its applicability has been tested on the Ben Ohau Range, New Zealand, where gradients of glacial erosion is recorded into the hypsometry of the modern topography. This test has demonstrated the high sensitivity of \( H_{kr} \) to spatial variations of glacial erosion. A numerical analysis helped to further constrain the use of \( H_{kr} \). Applying the hypsometric analysis to a region which includes the European Alps and the northern Apennines one can obtain insights into the major regional morphologic gradients induced by glaciation. The correspondence between the concentration of surface area and the LGM ELAs suggest the prevalence of a glacial buzz-saw effect as a consequence of glacial erosion. It
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is worth noting, however, that the study of spatially variable glacial imprint on the present-day topography of mountain ranges is limited by several factors such as, for example, post-glacial sediments that hide the actual shape of the glacially impacted bedrock. In addition, the contribution of tectonics, lithology, hillslope processes and precipitation rates in defining the hypsometry of a landscape are still largely unknown. Thus, results of such morphometric analyses must be treated with caution.

- Inverting channel head elevation data to resolve for pre-glacial channel steepness, $k_s$, enabled the reconstruction of the pre-glacial topography of the Alps. Spatial variations of $k_s$ highlight a region of high values in the western Alps which cuts across lithologic and tectonic province boundaries. The lack of evidence for young tectonic activity or lower precipitation rates restricted to this area suggest a potential change in regional isostatic support of the western Alps. For example, it has been suggested that this region experienced recent uplift by detachment of the European slab (e.g., Lippitsch et al., 2003). This region also exhibits younger thermochronologic ages, which indicates higher exhumation rates over the last few Ma (e.g., Vernon et al., 2008). The inferred pattern of pre-glacial $k_s$ across the rest of the range appears to be related to rock erodibility. In general, high values of pre-glacial $k_s$ are associated with harder rocks of the basement and metamorphic nappes, while low values correspond to weaker Alpine flysch deposits. The model of the pre-glacial topography of the Alps suggests that most glacial erosion has occurred in the lower and peripheral parts of the Alpine catchments, below the long term ELA. According to our preferred model, glacial erosion decreased mean elevation from 1375±641 m to 1310±720 m and increased valley scale topographic relief across the entire range, up to a factor of two. The method I developed can be applied to reconstruct the pre-glacial topography of any alpine mountain range and will provide information on pre-glacial tectonic and climatic conditioning of mountain ranges.

- Simulations of glacial erosion on the pre-glacial and present-day landscape suggest that the effects of glaciation are more complex than a simple “buzz-saw” erosion or increase of topographic relief. Glacial conditioning of a landscape evolves
through time, also in relation with the evolution of the landscape itself. Patterns of glacial erosion on the pre-glacial and present-day topography of the Rhône valley suggest an initial increase of topographic relief followed by enhanced erosion at higher elevations during late stages of glaciation. Valley over-deepening and enhanced glacial erosion at (or somewhat above) the long term ELA of the Rhône valley are linked by a headward propagation of glacial erosion through time. Such a headward propagation of glacial erosion is capable of matching evidences for a glacially induced increase of topographic relief (Valla et al., 2011) within the Rhône valley as well as increased glacial erosion in correspondence of the long term ELA (Sternai et al., 2011). In absence of any tectonic component of rock uplift, additional cooling trends superimposed over glacial cycles or an increase of the amplitude of climate oscillations to compensate for the decrease of glaciers’ accumulation area induced by glacial erosion, glacial sediment yield decreases during successive glacial cycles. If climate oscillations are constant through time and no additional, long term cooling trends are included, a tectonic component of rock uplift of 0.5 mm yr$^{-1}$ is sufficient to increase the production of sediments throughout glaciation, in agreement with measurements of sediment yields from the Alps (Hay et al., 1992; Kuhlemann, 2000; Kuhlemann et al., 2001). A shift of the period of climate oscillations from 40 kyr to 100 kyr can enhance the sediment yield by extending the time spans of glacial erosion and promoting new transient conditions. This mechanism or a change in regional isostatic support of the Western Alps (Lippitsch et al., 2003) are likely factors to explain the discrepancy between the age of the earliest glacial deposits in the northern foreland (Penck and Brückner, 1909; Schlüchter and Kelly, 2000; Häuselmann et al., 2007)) and in the Po plain (Muttoni et al., 2003) as well as geochronometric datasets suggesting an increase of valley deepening by glacial erosion at $\sim 1$ Myr (Häuselmann et al., 2007; Valla et al., 2011).
Appendix A

Glacial hydrology and erosion pattern: a mechanism for carving glacial valleys


A.1 Introduction

Although it seems obvious to the casual observer that glaciers erode, transport and deposit sediments, little is still known about what processes exactly control glacial erosion. Most glacial erosion models assume that erosion rate is proportional to ice-sliding velocity (e.g., Hallet, 1979; Oerlemans, 1984; Humphrey and Raymond, 1994; Hallet, 1996; Herman and Braun, 2008; Hildes et al., 2004; Egholm et al., 2009; Pelletier et al., 2010; Egholm et al., 2011). Using such a law, glacial erosion models have shown their abilities to reproduce typical glacial landscape features such as U-shape or hanging valleys (Harbor et al., 1988; Amundson and Iverson, 2006), but also suggest glacial erosion to be most efficient around and above the mean Equilibrium Line Altitude (ELA) (e.g., Anderson et al., 2006; Tomkin and Braun, 2002; Egholm et al., 2009). Such a mechanism is corroborated by large
scale topographic analyses showing that mean mountain elevation correlates with climate-controlled gradient in snow-line altitude (e.g., Penck, 1905; Broecker and Denton, 1989b; Porter, 1989; Brozovic et al., 1997; Mitchell and Montgomery, 2006; Egholm et al., 2009; Anders et al., 2010; Pedersen et al., 2010; Sternai et al., 2011; Champagnac et al., 2011).

Some of these models have also suggested that overdeepenings (i.e., a downglacier increase in elevation along the longitudinal profile) can result from an increase in ice discharge immediately below a tributary junction (MacGregor et al., 2000) or from the product of a positive feedback between topography and glacial erosion (Oerlemans, 1984; Kessler et al., 2008). Overdeepenings are prominent features of basins and valleys formed by glaciers and ice caps. They are typically found in high altitude glacial cirques (Hooke, 1991), but are also very common at low altitudes in the form of fjords, in-filled overdeepened glacial troughs or glacial lakes (e.g., Penck, 1905; Nesje and Whillans, 1994; Preusser et al., 2010; Durst-Stucki et al., 2010; Jordan, 2010). These low altitude features represent some of the most conspicuous products of erosion observed on Earth (e.g., Stern et al., 2005). They are found in most glaciated mountain ranges like the European Alps, Alaska, British Columbia or the Southern Alps of New Zealand as well as fjord-land margins of Patagonia, Greenland, Scandinavia or Antartica and testify to large amounts of erosion well below the altitude of river base level (i.e., the lowest point to which the river can flow), and locally below last glacial maximum (LGM) sea level. In Figure 1, we show a series of longitudinal profiles along glacial valleys at various locations in Norway, Switzerland and New Zealand. In these classical examples, intense erosion is required towards the downglacier reaches, within the ablation zone where water due to melting abounds. Such observations therefore led some authors to emphasize on the role of water either by suggesting that bedrock erosion depends primarily on basal water discharge (Alley et al., 1997; Durst-Stucki et al., 2010; Jordan, 2010) or by postulating that the role of subglacial water is to modulate the effective pressure (i.e., the difference between the water pressure and the ice overburden pressure) below the glacier and in turn promote erosion (Boulton, 1974; Hooke, 1991; Boulton, 1996; Hildes et al., 2004; Cohen et al., 2006).

In an attempt to represent the important role of water on glacial erosion, we incorporate a subglacial hydrology component into a glacial erosion model that
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is based on a sliding rule. We investigate how subglacial hydrology affects the patterns of glacial erosion by developing a model that includes three main components: (1) ice flow, (2) glacial erosion and (3) subglacial hydrology. This results in a fully coupled non-linear system of equations that we solve numerically. Our main objective is to investigate the influence of hydrology on the basal effective pressure, which is inversely proportional to sliding speed and hence erosion rates. Our modeling results suggest that this effect might have a first order control on erosion patterns. Moreover, it enables us to reconcile two contradictory observations: high erosion around the ELA and the formation of large overdeepenings towards low altitude.

A.2 Coupled ice dynamics, subglacial hydrology and erosion model

In this section, we present the different components of the model. We start by describing the ice model and then focus on subglacial hydrology and glacial erosion components. The approach we follow is similar to that presented in Hildes et al. (2004).

A.2.1 Ice model

The ice thickness, \(h\), is determined by solving the equation of mass conservation of ice,

\[
\frac{\partial h}{\partial t} = \nabla \cdot Q + M,
\]

where \(Q\) is the ice flux, \(t\) the time and \(M\) the mass balance that includes accumulation and ablation of ice. We only model temperate ice and assume simply that \(M\) is a linear function of the mean ambient temperature \(T_s\) (e.g., Meier et al., 1971; Herman and Braun, 2008), itself scaling linearly with elevation (i.e., \(M = -\gamma T_s\), where \(T_s = T_0 - \lambda (h + z)\), where \(T_0\) can be varied to simulate glacial cycles). Maximum rates (\(M_{\text{max}}\)) of ice accumulation and ablation (\(M_{\text{min}}\)) are also prescribed in order to stay within realistic bounds. \(Q\) is equal to the integral of ice velocity over
A.2 Coupled ice dynamics, subglacial hydrology and erosion model

Figure A.1: Examples of longitudinal profiles: (a) Sonje Fjord, Norway (Nesje and Whillans, 1994), (b) Rhone Valley, Switzerland (Jaboyedoff and Derron, 2005); (c) Whataroa River, New Zealand (Davey, 2010). The insets depict the location of the extracted profile.
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\( h \), where the velocity is the sum of the deformation velocity, \( u_d \), and the sliding velocity, \( u_s \), at the ice-bedrock interface. Deformation velocity is computed assuming Glen’s flow (e.g., Paterson, 1994). The approximations used for the velocity calculations are described in section A.3.1. While the fundamental law for sliding is still highly debated in glaciology (e.g., Schoof, 2005; Giagliardini et al., 2007), here we use an empirical sliding rule that is often found in numerical models (e.g., Bindschadler, 1983; van der Veen, 1987; Paterson, 1994; Pattyn, 1996; Muer and Vincent, 2003; Tomkin and Braun, 2002; Herman and Braun, 2008; Egholm et al., 2009)

\[
 u_s = \frac{B_s \tau_b^m}{p_e}, \tag{A.2}
\]

where \( \tau_b \) is the ice basal shear stress (which is to a first order proportional to ice thickness and ice slope surface), \( B_s \) a sliding constant, \( m \) an exponent between 1 and 3 and \( p_e \) the effective pressure (i.e., \( p_i - p_w \) where \( p_i \) is the ice overburden pressure and \( p_w \) the water pressure). Ultimately, the important part is the fact that water pressure controls the coupling between the ice and ground surface: low effective pressure weakens the ice-bed contact and permits faster sliding.

A.2.2 Subglacial hydrology

Several subglacial hydrology models can be found in the literature (e.g., Kessler and Anderson, 2004; MacGregor et al., 2008; Pimentel and Flowers, 2011; Flowers and Clark, 2002; Schoof, 2010). We adopt here the model of Flowers and Clark (2002) who proposed a complex glacial hydrologic model made of four components: surface ablation and runoff, englacial storage and transport, subglacial flow in a macroporous subglacial sheet and subsurface groundwater flow. In our model, we only consider flow in a porous subglacial sheet. While subsurface aquifers are important reservoirs for subglacial drainage due to their ability to delay and dampen subglacial pressures, they can only store a fraction of subglacial meltwater (Lemieux et al., 2008). This is even more true for crystalline and metamorphic rocks found such as the Alps which have low porosity and hydraulic conductivity. We only consider subglacial flow, which is the most important hydrologic component for sliding and therefore erosion. This assumption seems sensible in
the ablation area, where most of the melt occurs, because the size and density of crevasses are large enough that water can reach the bed easily. Finally, we do not consider the effects englacial meltwater may have on the ice rheology.

Following this model, the water thickness, $h_w$, is computed as follows

$$\frac{\partial h_w}{\partial t} = \dot{b} + \dot{b}_b - \nabla \cdot Q_w,$$  \hspace{1cm} (A.3)

where $Q_w$ the water discharge and $\dot{b}$ and $\dot{b}_b$ are source terms. It is clear that water within a glacier can have various origins (e.g., ice-snow melts, rainfall, run-off from hillslopes, geothermal heat flux, frictional heating or release of stored water (Paterson, 1994)). We simply assume that ablation is the primary source of water ($\dot{b}=|\min(M,0)|$) and add a background basal melt, $\dot{b}_b$. The flux of water $Q_w$ is defined as follows
A. Glacial hydrology and erosion pattern: a mechanism for carving glacial valleys

\[ Q_w = K_w h_w \nabla \psi, \]  
(A.4)

where \( K_w \) is the hydraulic conductivity and \( \psi \) the hydraulic head

\[ \psi = \frac{p_w}{\rho_w g} + z, \]  
(A.5)

where \( \rho_w \) is the water density, \( g \) the gravitational acceleration and \( z \) the bedrock elevation. The water pressure is computed following an empirical formula (Flowers and Clark, 2002)

\[ p_w = p_i \left( \frac{h_w}{h_c} \right)^\alpha, \]  
(A.6)

where \( p_i \) is the ice overburden pressure, \( h_c \) a critical thickness of the confined aquifer and \( \alpha \) an empirical exponent, for which Flowers and Clark (2002) suggested a value of 3.5. \( h_c \) represents either the averaged volume of a water-filled cavity system or the product of porosity and undilated layer thickness of a macroporous till sheet separating the ice and bed (Pimentel and Flowers, 2011). The main objective with this formulation is to include a nonlinear increase in water pressure with water thickness.

Finally, it is generally well-accepted that water flows either within a slow (i.e., usually poorly developed linked cavity system (Kamb, 1987; Fountain and Walder, 1998)) or fast drainage system (i.e., N-channels or R-channels (Nye, 1973; Rothlisberger, 1972; Fountain and Walder, 1998)). Following Flowers and Clark (2002), we do not include an explicit formulation of the channels but instead simulate the transition from slow to fast drainages by dynamically adjusting the hydraulic conductivity as a function of \( h_w \),

\[ \log(K_w) = \frac{1}{\pi} (\log(K_{\text{max}}) - \log(K_{\text{min}})) \cdot \arctan \left( k_a \left( \frac{h_w}{h_{\text{crit}}} - k_b \right) \right) + \frac{1}{2} (\log(K_{\text{max}}) + \log(K_{\text{min}})), \]  
(A.7)

where \( K_{\text{max}}, K_{\text{min}}, k_a \) and \( k_b \) are arbitrarily chosen (Flowers and Clark, 2002).

Finally, it is worth mentioning that the super-cooling effects, as water decom-
presses too quickly as it ascends up a reverse slope, are not explicitly accounted for (Alley et al., 2003).

A.2.3 Erosion model

Glacial erosion processes are usually acknowledged to be dominated by abrasion, quarrying or subglacial streams. Even though the latter may play an important role (e.g., Alley et al., 1997; Durst-Stucki et al., 2010; Jordan, 2010), we assume that abrasion and quarrying are the dominant processes and are proportional to sliding velocity, as suggested previously and commonly adopted in landscape evolution models (e.g., Hallet, 1979; Hooke, 1991; Hallet, 1996; MacGregor et al., 2000; Tomkin and Braun, 2002; Anderson et al., 2006; Herman and Braun, 2008; Pelletier et al., 2010; Egholm et al., 2009, 2011),

$$\frac{\partial z}{\partial t} = K_g u_s^l,$$

(A.8)

with $K_g$ is an erosion constant, $l$ an erosion coefficient ranging from 1 to 4 (Harbor et al., 1988). Erosion rates are multiplied by the cosine of the bedrock slope to avoid runway effects (i.e., unrealistic erosion rates when the bedrock slope becomes too large) and we assume that all debris are efficiently transported out of system, as indicated by the centimeter scale debris thickness observed near the ice-bedrock contact (e.g., Boulton, 1970; Humium, 1996).

A.3 Numerical models

In this section, we first outline the numerical methods we adopt and then report a series of simulations, starting with the hydrology alone followed by the coupled model. For these latter simulations, we first show 1D longitudinal profiles and then describe results for a 2D planview model. This strategy was designed to derive an optimal model parameterization; which can in turn be used to guide to more computationally expensive two-dimensional models.
A. Glacial hydrology and erosion pattern: a mechanism for carving glacial valleys

A.3.1 Numerical methods

Subglacial hydrology (equation A.3) is solved using the finite volume method. Time stepping is treated explicitly, using a Picard iterative scheme to deal with the non-linearity.

For the 1D profiles, we use a two-dimensional model of ice dynamics for temperate ice in which ice deformation is solved using the hydrostatic approximation introduced by Blatter (1995). In contrast with the shallow ice approximation (Hutter, 1983), which is often used for glacial erosion models, longitudinal stresses are taken into account. The ice velocities are computed using the method described with great detail in Blatter (1995). The time integration (equation A.1) is computed by first integrating the ice flux vertically and then computing its divergence using the finite volume method. Time stepping is treated explicitly using a Lax-Wendroff scheme (Lax and Wendroff, 1960).

For the 2D planview models, we use a modified version of the numerical model reported in Tomkin and Braun (2002) and Herman and Braun (2008), which is based on the shallow ice approximation (Hutter, 1983). Both the ice and subglacial hydrology equations (equations A.1 and A.3) are solved explicitly, also using a Picard iterative scheme.

Finally, it is important to emphasize that there are little constraints on the various model parameters. We chose the parameters for the hydrology model following Flowers and Clark (2002) and Pimentel et al. (2010). Our goal is not to test this choice, which is further explored in Pimentel et al. (2010) and Pimentel and Flowers (2011). Furthermore, trade-offs between parameters such as sliding and erosion constants exist, which also makes the choice of parameters difficult. Therefore, the actual parameter values must be treated with caution. However, these can be tuned to mimic plausible erosion rates that are consistent with rates observed in nature using methods such as cosmogenic nuclide dating (e.g., Fabel et al., 2004), $^4\text{He}/^3\text{He}$ thermochronometry (Shuster et al., 2005) or OSL-thermochronology (Herman et al., 2010).
A.3 Numerical models

Figure A.3: Model 1, with model parameters shown in Table A.2. The top part represents the glacier and bedrock evolutions in time. The middle part is the erosion rates during the simulation and integrated erosion at the end of the run. The bottom part corresponds to the hydraulic conductivity. Results are plotted every 30 ka. The total run time is 90 ka.

A.3.2 Subglacial hydrology model alone

In Figure A.2a, we present the result of a simulation in which we impose a constant ice geometry that has a Gaussian shape and lies on a flat bed. We also prescribe a constant water input (\( \dot{b} = 10 \text{ m/a} \)) over the entire domain. We set \( h_w \) equal to zero as initial condition and run the model until it reaches steady state (see model parameters in Table A.1). The objective of this model is to assess the time response of the hydrologic model only.

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As one might expect, this simulation leads to a higher water thickness towards the edges of the ice cap. This is because water flux \((Q_w)\) is driven by the gradient in water potential \((p_w)\), which is itself inversely proportional to the ice pressure and water thickness (equation A.6). Since the ice thickness is higher in the middle, water is simply driven towards the thinner edges.

We report the transient solution in Figure A.2b, where we plot the maximum water thickness \((h_w)\) as a function of time. It shows that the response time is proportional to the water input, \(\dot{b}\), which is consistent with a diffusive model. Although it may be longer when \(\dot{b}\) is low (i.e., <1 m/a), it remains within the time scale of a melt season in most cases (i.e., less than three months).

Water input is also known to vary on a daily basis during the summer season, which has an important effect on sliding velocities (e.g., Iken and Bindschalder, 1986; Bartholomaus et al., 2007). To test how such process might influence our model, we modulate \(\dot{b}\) by multiplying it by \(\sin(2\pi \text{day})\). The main effect is to increase the length of time to reach steady state. This is simply because, first, the total water input is smaller compared to when water input is constant and, second, that the response time of the system is longer than a day.

Finally, because we are mainly interested in the effects such model would have on erosion and sliding, we also plot the minimum normalized effective pressure (i.e., \(\frac{h_n - p_w}{p_i}\)) in Figure A.2b. It shows that low effective pressure in our model is reached before the hydrologic model reaches steady state. This is because of the non-linearity imposed in equation 2.6. In turn, it suggests that the response time in sliding and erosion to water inputs remains short (i.e., less than 1 month).

A.3.3 Coupled model, 1D longitudinal profile

We show in Figure A.3 and A.4 the results of simulations that are run for 90 ka, which is a timescale comparable to a Quaternary glacial cycle. The parameters were scaled (Table A.2) to obtain maximum erosion rates of the order of 10 mm/a and a mean erosion rate across the profile of the order of 1 mm/a. The climate is kept constant in all simulations. This is obviously an oversimplification, but it enables us to explore the hydrologic effects on erosion patterns. The location of the ELA is set at 1800 m above present-day sea level. We assume that the long
A.3 Numerical models

profile started off as a 35 km long fluvial profile with a fluvial relief going from 500 to 3472 m, as an initial condition. Our goal is to (1) illustrate the influence of glacial hydrology on the erosion patterns, (2) explore the importance of the erosion law exponent \( l \) in equation A.8, (3) investigate the effects of varying the exponent relating \( h_w, p_i \) and \( p_{w} \) for the hydrologic model \( \alpha \) in equation A.6, (4) estimate the relevance of hydraulic conductivity adjustments (equation A.7) and (5) assess the importance of seasonal water input on erosion patterns.

As expected, the main effect of including subglacial hydrology is to reduce...
A. Glacial hydrology and erosion pattern: a mechanism for carving glacial valleys

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \dot{b} )</td>
<td>10 m/yr</td>
<td>water input</td>
</tr>
<tr>
<td>( \dot{b}_b )</td>
<td>0 m/yr</td>
<td>background water input</td>
</tr>
<tr>
<td>( h_c )</td>
<td>0.1 m</td>
<td>critical water thickness</td>
</tr>
<tr>
<td>( K_{min} )</td>
<td>( 1 \times 10^3 )</td>
<td>minimum hydraulic conductivity</td>
</tr>
<tr>
<td>( K_{max} )</td>
<td>( 1 \times 10^6 )</td>
<td>maximum hydraulic conductivity</td>
</tr>
<tr>
<td>( k_a )</td>
<td>15.0</td>
<td>hydraulic conductivity adjustments constant</td>
</tr>
<tr>
<td>( k_b )</td>
<td>0.85</td>
<td>hydraulic conductivity adjustments constant</td>
</tr>
</tbody>
</table>

Table A.1: Parameter values used for hydrologic model alone (Figure A.2).

the effective pressure in the ablation area and in turn increase sliding, as clearly shown by the simulation. The bottom part of Figure A.3 (Model 1) shows that the hydraulic conductivity remains low in the upper part of the glacier, where water input is low, and increases dramatically within the ablation area. As the model progresses forward in time, the erosion rate is maximized below the ELA to ultimately lead to a local overdeepening. A wave of erosion traveling upstream the profile is observed as the model evolves. Interestingly, a positive feedback appears: the velocity and, thus, erosion rates increase, in turn augmenting the local gradient and the sliding velocity itself. These results suggest that erosion rates are expected to gradually increase as the topography switches from a fluvial to a glacial landscape. However, the accumulation area keeps being reduced as the glacier erodes, in turn limiting glacial fluxes and average erosion rates. Finally, it is interesting to note that the total erosion (i.e., integrated over the total run (green line in Figure A.3a)) is maximized around the location of the ELA but its distribution is different compared to the patterns of erosion rates at particular times 30 ka apart (blue line in Figure A.3a), which are maximized down in the ablation area.

In Figure A.4a (Model 2), we set \( l \) in equation A.8 equal to 2. Note that a direct comparison with the previous model is difficult because \( K_g \) must be adjusted to obtain similar erosion rates. In this case, the main effect is to change the shape of the ‘erosion rate function’, with erosion being more localized around its maximum value. Surprisingly, the overdeepening that is produced is not as deep as observed
Table A.2: Parameter values used to compute ice thickness in the 1D longitudinal profile, Model 1 to 5 (note the ice is model 2D (Blatter, 1995)).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$B$</td>
<td>$6 \times 10^{-24}$ Pa$^{-3}$ s$^{-1}$</td>
<td>flow law parameter</td>
</tr>
<tr>
<td>$B_s$</td>
<td>$7 \times 10^{-16}$ Pa$^{-3}$ s$^{-1}$ m$^{-2}$</td>
<td>sliding law parameter</td>
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<td>$n$</td>
<td>3</td>
<td>Glen’s flow exponent</td>
</tr>
<tr>
<td>$\rho$</td>
<td>910 kg m$^{-3}$</td>
<td>density of ice</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>1000 kg m$^{-3}$</td>
<td>density of water</td>
</tr>
<tr>
<td>$g$</td>
<td>9.81 m s$^{-2}$</td>
<td>gravitational acceleration</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>$4.5 \times 10^{-3}$ °Cm$^{-1}$</td>
<td>lapse rate</td>
</tr>
<tr>
<td>$M_{\text{max}}$</td>
<td>2 m/yr</td>
<td>maximum accumulation rate</td>
</tr>
<tr>
<td>$M_{\text{min}}$</td>
<td>-20 m/yr</td>
<td>maximum ablation rate</td>
</tr>
<tr>
<td>$\dot{b}_b$</td>
<td>$10^{-3}$ m/yr</td>
<td>background water input</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>2.0</td>
<td>arbitrary constant</td>
</tr>
<tr>
<td>$h_c$</td>
<td>0.1 m</td>
<td>critical water thickness</td>
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<td>$K_{\text{min}}$</td>
<td>$3.15 \times 10^3$</td>
<td>minimum hydraulic conductivity</td>
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<tr>
<td>$K_{\text{max}}$</td>
<td>$8.50 \times 10^6$</td>
<td>maximum hydraulic conductivity</td>
</tr>
<tr>
<td>$k_a$</td>
<td>15.0</td>
<td>hydraulic conductivity adjustments constant</td>
</tr>
<tr>
<td>$k_b$</td>
<td>0.85</td>
<td>hydraulic conductivity adjustments constant</td>
</tr>
<tr>
<td>$K_g$</td>
<td>$2.0 \times 10^{-4}$ - $3.0 \times 10^{-6}$</td>
<td>glacial erosion constant</td>
</tr>
<tr>
<td>$l$</td>
<td>1-2</td>
<td>glacial erosion exponent</td>
</tr>
<tr>
<td>$T_0$</td>
<td>8 °C</td>
<td>fixed mean temperature</td>
</tr>
</tbody>
</table>

In Figure A.4b (Model 3), we reduce $\alpha$ in equation A.6 to 1. Similarly to Model 2, it has a limited impact on the patterns of glacial erosion. Therefore, together with the findings reported in Figure A.3, it appears that the effect of subglacial hydrology remains relatively insensitive to the choice of $l$ and $\alpha$.

In Figure A.4c (Model 4), we only allow the hydraulic conductivity to increase by one order of magnitude instead of three. Such a simulation is equivalent to a glacier where subglacial hydrology is entirely dominated by a slow drainage system. This simulation leads to a rather uniform erosion pattern, which is similar to a
A. Glacial hydrology and erosion pattern: a mechanism for carving glacial valleys

situation where subglacial hydrology is not accounted for.

Our primary goal is to assess the averaged effects water has on the long-term patterns of erosion (i.e., tens to hundreds of thousands of years). However, short time variations in water input have an effect on sliding velocities (e.g., Bartholomaeus et al., 2007; Joughin et al., 2008). Therefore, we present a model in which seasonal variations are included (Figure A.4d, Model 5) by assuming that water reaches the bed only during one month. After appropriate tuning of the erosion constant \(K_g\) because the amount of time when the effective pressure is low is shorter, we obtain rates and patterns of erosion that are comparable to the other models. The explanation is found in the time response of the hydrologic system that is short compared to the time scale of erosion.

A.3.4 Coupled model, 2D plan-view topography

We apply the 2D plan-view model using actual topography. The ice model we use is reported in detail in Herman and Braun (2008). We chose to use the western European Alps as a natural example because the depth of the overdeepenings along the Rhone valley is constrained by geophysical and empirical data (Jaboyedoff and Derron, 2005) (Figure A.1 and Figure A.5). It is not our purpose to explain the geomorphology of this area (which would require including more complexity, such as more detailed climatic, lithologic or tectonic components and is beyond the scope of the present study) but rather to provide a first order comparison for our modeling results. In Figure A.5, we show the evolution of topography in a forward simulation that lasts for 1 Ma and in which we impose a series of ten glacial cycles, each of 100 ka duration, with an ELA that oscillates following simple sinusoidal variations between 2800 and 1750 m (van der Beek and Bourbon, 2008). The initial conditions consist of a reconstructed pre-glacial topography, which is simply obtained by replacing the in-fill gravels with hard rocks. The parameters used in the models are reported in Table A.3. The significant difference between our initial condition and the present-day bedrock geometry is highlighted by comparing Figure A.5a with Figure A.5b, while the evolving modeled topography and predicted amounts of erosion are depicted in Figure A.5c to A.5f.

Recent studies in the Alps have suggested that most glacial depressions were
A.3 Numerical models

Figure A.5: 2D planview erosion model results. The model is run for 1 Ma, with temperature oscillations of 5°C (corresponding to fluctuations of the ELA between 1750 and 2800 m) with a period of 100 ka (which corresponds to similar values observed in the Alps). (a) Initial topography used in the model. Black lines are longitudinal profiles presented in Figure A.7; (b) Bedrock topography model (Jaboyedoff and Derron, 2005) highlighting the geometry of the overdeepened Rhone valley (note that depth of the overdeepening is only available for the Rhone valley and not for its tributaries); (c) and (d) Modeled topography at time $t_1$ and $t_2$ ($t_1=100$ ka and $t_2=1000$ ka); (e) Difference between topography shown in (c) and (a); (f) Difference between topography shown in (d) and (a).

formed progressively over several glacial cycles, mainly based on lithostratigraphic and dating record (Preusser et al., 2010; Durst-Stucki et al., 2010; Jordan, 2010), that occurred over the last $\sim$1 Ma (Muttoni et al., 2003; Haeuselmann et al., 2007; Valla et al., 2011, 2012). The results presented in Figure A.5 illustrate how one
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<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$B$</td>
<td>$6 \times 10^{-24} \text{ Pa}^{-3} \text{ s}^{-1}$</td>
<td>flow law parameter</td>
</tr>
<tr>
<td>$B_s$</td>
<td>$5 \times 10^{-15} \text{ Pa}^{-3} \text{ m}^{-2}$</td>
<td>sliding law parameter</td>
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<td>$n$</td>
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<td>Glen’s flow exponent</td>
</tr>
<tr>
<td>$\rho$</td>
<td>$910 \text{ kg m}^{-3}$</td>
<td>density of ice</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>$1000 \text{ kg m}^{-3}$</td>
<td>density of water</td>
</tr>
<tr>
<td>$g$</td>
<td>$9.81 \text{ m s}^{-2}$</td>
<td>gravitational acceleration</td>
</tr>
<tr>
<td>$k_c$</td>
<td>500 m</td>
<td>constriction constant</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>$4.5 \times 10^{-3} \degree \text{ C m}^{-1}$</td>
<td>lapse rate</td>
</tr>
<tr>
<td>$M_{\text{max}}$</td>
<td>2 m/yr</td>
<td>maximum accumulation rate</td>
</tr>
<tr>
<td>$M_{\text{min}}$</td>
<td>-20 m/yr</td>
<td>maximum ablation rate</td>
</tr>
<tr>
<td>$b_b$</td>
<td>$10^{-3} \text{ m/yr}$</td>
<td>background water input</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>2.0</td>
<td>arbitrary constant</td>
</tr>
<tr>
<td>$h_c$</td>
<td>0.1 m</td>
<td>critical water thickness</td>
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<tr>
<td>$K_{\text{min}}$</td>
<td>$1.0 \times 10^3$</td>
<td>minimum hydraulic conductivity</td>
</tr>
<tr>
<td>$K_{\text{max}}$</td>
<td>$1.0 \times 10^6$</td>
<td>maximum hydraulic conductivity</td>
</tr>
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</tr>
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<td>$K_g$</td>
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<td>glacial erosion constant</td>
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<tr>
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<td>glacial erosion exponent</td>
</tr>
<tr>
<td>$T_0$</td>
<td>8-13 $\degree \text{C}$</td>
<td>temperature variations</td>
</tr>
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Table A.3: Parameter values used to compute ice thickness in the 2D plan view simulations (Herman and Braun, 2008).

can gradually produce well-developed overdeepenings in the lower reaches of the glaciers, where effective pressure is low and glacial erosion is high. As the simulation evolves in time, the overdeepenings propagate upward and, at last, erosion becomes more intense within the higher reaches of the catchments. Our model results are therefore consistent with the commonly accepted idea that erosion is significant around the ELA, but they also predict large amounts of erosion far below the ELA. This result is further demonstrated by performing a hypsometric analysis of the topography (i.e., the frequency distribution of elevations) and
comparing these results to a situation in which subglacial hydrology is turned off. In both cases, the results exhibit significant erosion near the mean location of the ELA (Figure A.6a and A.6b), but enhanced erosion at low altitude is observed when subglacial hydrology is included (Figure A.6b), leading to a bimodal distribution of erosion. This is an important result because it enables us to reconcile two apparently contradictory observations: (1) glacial erosion plays a major role on controlling mean mountain height topography and (2) glaciers are also capable of forming deep erosional features such as glacial lakes and fjords.

![Bi-variate hypsometry: the y-axis represents the initial altitude and the x-axis the thickness of material that was eroded during the simulation. The color scheme represents the frequency of pixels for the given initial elevation and erosion. (a) Model without subglacial hydrology. (b) Model that includes subglacial hydrology.](image)

Figure A.6: Bi-variate hypsometry: the y-axis represents the initial altitude and the x-axis the thickness of material that was eroded during the simulation. The color scheme represents the frequency of pixels for the given initial elevation and erosion. (a) Model without subglacial hydrology. (b) Model that includes subglacial hydrology.

We extract in Figure A.7 a series of five longitudinal profiles along the Rhone Valley (Profile 1 in Figure A.5a) and some of its tributaries (Profile 2 to 5 in Figure A.5a). The profiles shown in Figure A.7a to A.7C are model solutions, not the present-day topography. These plots further emphasize the remarkable amount of erosion predicted in the downstream part of the profiles, with more
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than a kilometer of rocks being eroded within the ablation zone of the ice cap during glacial maximum periods. Our model also predicts the formation of hanging valleys between the main trunk and its tributaries. Additionally, it is worth noting that the overdeepenings do not necessarily match with confluences, which has been previously suggested by the 1D numerical models (MacGregor et al., 2000) but is not always observed in the field (e.g., Preusser et al., 2010). The larger amounts of erosion (i.e., several hundreds of meters) predicted within the glacial cirques are primarily due to higher local gradients, which in turn cause faster sliding velocities and erosion rates.

In Figure A.7c, we also present the evolution of the main valley profiles during a simulation in which a cold climate is imposed, i.e., an ELA fixed at 1750 m (Figure A.7c). Because the ice surface slope is low in the core of the ice cap and there is little water present at the ice-bedrock interface, erosion is only acting within the ablation zone. This, in turn, leads to a much larger overdeepened trough at the snout of the ice cap. Erosion slows down when the overdeepened zone is deep enough that ablation increases substantially, ultimately causing a reduction of the ice extent. This situation could be similar to high latitude fjord-land regions. For instance, Thomson et al. (2010) recently inferred low erosion rates from low temperature thermochronological data and provided support for the idea that frozen-based glaciers can protect mountain peaks (e.g., Griffiths, 1952; Oerlemans, 1984; Stroeven et al., 2002; Tomkin and Braun, 2002; Stern et al., 2005). The model presented here suggests that having a large ice cap would suffice to have limited erosion rates in the central part of the range without requiring the ice to be frozen to the bedrock.

A.4 Discussion and conclusions

Subglacial hydrology affects erosion rates in a number of ways. Vigorous water discharges may strip bedrock of debris, allowing underlying rock to be eroded more rapidly by sliding ice than if rock remained sheltered by debris. Variable water discharge may lead to water and ice pressure variations on the bed that fatigue and crack bedrock, thereby hastening quarrying (e.g., Hallet, 1996; Iverson, 2002; Cohen et al., 2006). Hydraulic potential gradients in the bed may exert
A.4 Discussion and conclusions

body forces on basal sediments and cracked bedrock, thereby helping to mobilize them (e.g., Alley et al., 1997). Similarly, there are geomorphic observations such as fluvial streamlines, glacial potholes or tunnel valleys, as documented in detail for the Alps (Durst-Stucki et al., 2010; Jordan, 2010), that imply a partial, or complete, decoupling between ice and meltwater erosion. We only investigated here the effect of basal hydrology on setting the basal effective pressure, which is inversely proportional to sliding speed and hence erosion rates, using a diffusive model. The main consequence of including subglacial hydrology is to reduce the effective pressure in the ablation area, which in turn leads to increased sliding and erosion in the downglacier reaches. Although we explore only one aspect of including subglacial hydrology, this may bear important implications for our understanding of glacial erosion processes in Alpine context. It provides us with a unifying explanation on how glaciers may control the mean elevation of mountain ranges while forming glacial lakes in the foreland.

The transition to subglacial channels is not explicitly simulated because of the adopted macroporous continuous hydrologic model. Consequently, the effective pressure remains low in the ablation area throughout most of the simulations, even though the hydraulic conductivity is increased by three orders of magnitudes with respect to its initial value. This is important because subglacial channels are thought to have a first order effect in reducing the water pressure in the ablation zone (e.g., Rothlisberger, 1972; Schoof, 2010). However, seasonal fluctuations of meltwater regimes have a strong control on channels dimension (Boulton et al., 2007) and sliding velocities (e.g., Iken and Bindschadler, 1986; Bartholomaus et al., 2007; Joughin et al., 2008; Pimentel et al., 2010; Schoof, 2010). Summer drives more subglacial meltwater than winter and drainage efficiency evolves accordingly. Therefore, it is likely that every year, there is a time period when drainages shift from slow to fast to discharge subglacial meltwater. During that time, subglacial water pressure is large and effective pressure is low, which promotes sliding and erosion. We postulate here that this is the critical period for erosion to occur. However, this does not need to be taken into account in the models because seasonal variations in water inputs have little effects on the long term patterns and rates of erosion, as we suggest in this study.

Qualitative comparison between model predictions and actual bed topography
A. Glacial hydrology and erosion pattern: a mechanism for carving glacial valleys

Figure A.7: Time evolution of the longitudinal profiles (location of these profiles is depicted by the black lines in Figure A.5). Each line represents the topographic evolution at the beginning of the simulation, after 100 ka, 500 ka and 1000 ka; (a) Same model results as the model presented in Figure A.5; Each color corresponds to a different profile (black = Profile 1; blue = Profile 2; magenta = Profile 3; green = Profile 4; red = Profile 5); Gray shading corresponds to the location of the ELA; (b) Same as (a) when the effect of subglacial hydrology is turned off; The dashed line corresponds to the end model to Profile 1 at the end of the model presented in (A); (c) Model results when climate is kept constant; (d) Comparison between model predictions from (a) and (b) bedrock topography inferred from empirical evidence (Jaboyedoff and Derron, 2005) and shown in Figure A.1.

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reconstructed from geophysical methods (Jaboyedoff and Derron, 2005) (Figure 4d) reveals that the modeled overdeepenings are similar, in terms of both location and depth, to those in the profiles reconstructed from empirical evidence. This is particularly true for the case in which subglacial hydrology is included and cyclic climate conditions are used. Our model results also predict patterns of erosion that display a definite clustering that evolves through time (i.e., that glacial erosion is more localized compared to fluvial erosion). The pattern of erosion is thus different for each stage in the climate evolution. This implies that (1) erosion by glaciers is not as uniform as by rivers and (2) local overdeepenings develop in different places at different time, depending on ELA location, local slope, basin hypsometry and the amount of ice accumulated upstream. This was already suggested by previous modeling studies (Herman and Braun, 2008), but these effects turn out to be even more pronounced when subglacial hydrology is incorporated. These results suggest that climatic oscillations control and rapidly modify erosion patterns, even in our simplistic paleo-climate scenario.

Finally, several studies have demonstrated that glacial erosion plays not only an important role on the large scale and Ma deformation patterns of mountain ranges (e.g., Koons, 1989; Meigs and Sauber, 2000; Tomkin and Roe, 2007), but also has the potential to modulate slip rates on localized fault structures at a ka time scale (Wu et al., 1999; Hetzel and Hampel, 2005). Our findings that the long-term distribution of erosion is bimodal and glacial erosion patterns can be clustered and evolve through time must have a strong influence on deformation patterns and rates in mountain ranges, at all spatial and temporal scales. In addition, our novel model provides a potential mechanism for explaining the formation of fjords in high latitude regions, which has remained unclear until today. Most fjords were formed on the edges of large ice caps, where water due to melting affects basal processes (e.g., Schoof, 2010) and thus erosion.
Appendix B

Reconstruction of Lateglacial stadia in the Matter and Saas Valleys, Canton Valais, Switzerland; implications for rock slope stability


B.1 Introduction

The landforms and processes of the central European Alps appear to be dominated by transient effects of glacial and periglacial processes. This is evident in the morphology of the deep “U”-shaped valleys, which derive their form from the effects of repeated Pleistocene glacial cycles, as well as ongoing hillslope and sedimentary processes which reflect changes in environmental conditions following the termination of the last glaciation. Dynamic periglacial landforms in the present-day landscape are particularly sensitive to ongoing effects of climate change. Under-
standing glacial and periglacial processes, and particularly their variability with
time, can therefore offer important insight into the response of Alpine landscapes
to a variety of environmental conditions. In this study, we combine field obser-
vations with cosmogenic nuclide dating and numerical modeling to unravel the
Lateglacial history of the Matter and Saas Valleys (Canton Valais, Switzerland).
In particular, we focus on a prominent sequence of moraines that delineate a no-
table ice re-advance into the trunk valleys, which also corresponds to a distinct
change in rock slope instability on the valley walls. Although morphological char-
acteristics suggest these moraines belong to the Egesen Alpine Lateglacial stadial,
the extent of the re-advance is greater than many Egesen glaciers mapped else-
where in the Alps, and positioning these moraines within the documented sequence
of Lateglacial stadia therefore requires further analysis. The Egesen ice advance
(representing the Younger Dryas cold phase) is unique in the Lateglacial record
as it followed abruptly from the Bølling/Allerød interstadial, the first significant
northern hemisphere shift to near-present-day climatic conditions since at least
the middle Würm glaciation (dated at \(\sim\)60-28 ka BP, see Doppes et al. (2011) and
references therein). This transition consisted of a significant glacial retreat, inter-
stadial, and re-advance within inner Alpine valleys, during which steep deglaciated
bedrock walls in the upper reaches were exposed for approximately 2000 years.

Although moraine deposits record the extent of temporary stadia during a
glacial termination event, the record is typically fragmented, and determining the
magnitude and timing of ice fluctuations following an event like the Last Glacial
Maximum (LGM) can be challenging. Further information on glacier fluctuations
during deglaciation can be derived from environmental conditions using climate
proxies, such as \(\delta^{18}O\) variations in ice cores (Johnsen et al., 2001) and pollen pre-
served in sediment deposits (Furrer et al., 1987). However, the advance and retreat
of individual glaciers is strongly dependant on a number of local factors particular
to the specific glacier system, including variations in climate, ice rheology, and
topography, which make assessing fluctuations from regional proxies difficult.

We combine morphological analysis and cosmogenic nuclide dating to describe
the spatial and temporal evolution of valley glaciers culminating in the deposi-
tion of a Lateglacial moraine sequence to moderate elevations within the Matter
and Saas Valleys. This stadial provides a reference point through which we are
able to calibrate a numerical model of Lateglacial ice extents based on the Shallow Ice Approximation (SIA, (Hutter, 1983)). The calibrated ice model does not replicate detailed glacial and climatic processes, but rather by applying realistic environmental and physical parameters represents a best fit of available climatic and morphological data, and provides insight into the dynamics of Lateglacial ice re-advances within the study region. Glacier mass balance calculations within the model are relatively robust, however, and we use evidence derived from estimates of Lateglacial equilibrium line altitude (ELA) depression in local cirque glaciers to correlate ELA depressions for model ice extents coinciding with mapped moraines in the study area to Lateglacial stadia elsewhere in the Alps.

B.2 Description of the study area

The Matter and Saas Valleys were major tributaries to the Rhone Glacier during the LGM. They are oriented sub-parallel and approximately north-south, and are about 10 km apart (Figure B.1). The geometries of the catchments are similar, however the Saas Valley lacks the extensive headwater region characteristic of the Matter Valley south of Zermatt, and as a result the area of the Matter Valley is almost twice that of the Saas (450 km$^2$ and 250 km$^2$, respectively). Valley floor elevations in both vary between 700 m and 2200 m, while the surrounding peaks are some of the highest in the Alps, reaching above 4500 m. The geomorphology of both catchments is dominated by the effects of glacial and periglacial processes; glacial landforms, polished bedrock surfaces, and till deposits are common in the main valleys and their tributaries, particularly at elevations above 2000 m. Although a more temperate climate close to the valley floor, combined with steep slopes (commonly over 40°) and numerous slope instabilities, has severely limited the preservation of glacial deposits at lower elevations.

B.2.1 Topography and geology

The study area is situated within two sub-zones of the Penninic nappes: the Grand-St.Bernard multi-nappe system (namely the Siviez-Mischabel nappe) and the Zermat-Saas Fee zone. The Siviez-Mischabel nappe includes primarily Permian-
B.2 Description of the study area

Carboniferous crystalline rocks such as ortho- and para-gneisses, mica-rich schists, and quartzites (Bussy et al., 1996; Escher et al., 1993). The Zermatt-Saas Fee zone is located in the headwaters of both valleys, and contains ultramafic rocks of ophiolitic origin (Bartolini et al., 2004). The lithological contrast between the two nappes is particularly useful for identifying glacially transported sediments, as serpenite boulders from the Zermatt-Saas Fee Zone in the upper catchments are usually highly-polished and striated as a result of glacial transport, and their characteristic dark green colour contrasts strongly against the quartz-rich gneisses of the Siviez-Mischabel nappe that dominate sedimentary deposits in the valleys. Otherwise, the ubiquity of ortho- and para-gneisses throughout most of the study area makes distinguishing moraine from other sedimentary deposits (particularly

Figure B.1: Geomorphological overview of the study region, including mapped Quaternary moraines, LGM trimlines, and LIA glacier extents. Selected cosmogenic $^{10}$Be and radiocarbon sample locations are also indicated.
B. Reconstruction of Lateglacial stadia in the Matter and Saas Valleys, Canton Valais, Switzerland; implications for rock slope stability

rockfall) difficult, since the deposits are commonly intermixed.

B.2.2 Lateglacial history

Alpine Lateglacial

The Last Glacial Maximum (dated at 24-21 ka BP (Schluchter, 1988, and references therein)) marks the peak of the late Würm glaciation, during which Alpine glaciers extended north to the Jura and beyond Lake Constance in northern Switzerland. This glaciation began at approximately 115 ka BP, and was characterized by an early advance, followed by an interstadial with temporarily ice-free Alpine forelands between 60 ka BP and 28 ka BP (Ivy-Ochs et al., 2006a; Schlüchter and Rothlisberger, 1995; Welten, 1982).

The transition from LGM to Holocene interglacial conditions occurred during the GS-2b warmer interval, beginning with an early lateglacial phase of ice decay around 21.1 ka BP which initiated the retreat of ice from the Alpine foreland (Figure B.2) (Ivy-Ochs et al., 2004; Reitner, 2007). Stratigraphy determined from high resolution seismic surveys of Lake Geneva records two re-advances of the Rhone glacier (Fiore et al., 2011) during this period, before its eventual “Lateglacial” retreat (Walker et al., 1999) of glaciers into the inner Alpine valleys. This recession back from the mountain front is currently poorly constrained in the Alps, though radiocarbon dates from a peat bog in the lower Traun Valley indicate that this eastern Alpine valley was ice free by approximately 18-19 ka BP (15.4±0.47 ka BP C) (van Husen, 1977). In the western Alps a compilation of radiocarbon dates in the Lake Geneva region suggests glacier retreat into the Rhone Valley occurred prior to 16.1±0.5 ka BP, though further work is required to determine the timing of deglaciation in this region (see Wohlfarth et al., 1994, and references therein).

This early recession was then followed by the collapse of the Alpine glacier system, which included a series of glacial re-advances in tributary Alpine valleys up until the end of the Egesen stadial at ∼11.6 ka BP (Ivy-Ochs et al., 2006b; Schluchter, 1988). These re-advances (or Lateglacial stadia) are designated from oldest to youngest: Gschnitz, Clavadel, Daun, Egesen, and Kartell (Penck and Brückner, 1909). Glacier fluctuations continued throughout the Holocene (Holzhauser, 2008; Holzhauser et al., 2005; Hormes et al., 2001), with extents varying between near
present-day, and a more extensive series of moraines 1-2 km down-valley which were most recently occupied by “Little Ice Age” (LIA) glaciers around 1850 AD. These distinctive outer (or LIA) moraines provide a useful reference to compare glacier extents during similar interstadial conditions.

Figure B.2: Compilation of dates determined from 10Be and radiocarbon sample sites discussed in the text. All dates are recalculated from original data using the methodology described in Section Sampling method and preparation. Filled markers indicate dated moraines representing deposition during glacial stadia, half-filled markers are from dated erratic boulders or bedrock surfaces, representing exposure during ice retreat. See Figure B.2 for locations of Randa, Saas Balen, and Saas Almagell sample locations. $\delta^{18}$O data from the GRIP ice core is plotted for reference, and indicates relative warm and cool intervals including GS-2b, the Oldest Dryas (OD), Bølling (B), Allerød (A), Younger Dryas (YD), and the Holocene (H).

Regional comparisons of glacial advance or retreat are commonly described in terms of their Equilibrium Line Altitude (ELA) depression relative to that of local LIA glaciers. The ELA is the elevation at which annual ablation and accumulation are balanced (Armstrong et al., 1973; Gross and Kerschner, 1977), and is best
determined by a series of direct mass balance measurements across the surface of a glacier throughout the course of a “representative” year. Such measurements are both time consuming and expensive, and as a result ELA has only been accurately measured for a few modern-day glaciers (<300, (see Zemp et al., 2009)). In order to perform regional or paleo-ELA reconstructions, a number of methods have been proposed that derive ELA estimates from remote measurements (see Osmaston, 2005, and references therein). The accumulation area ratio (AAR) is one of the most common methods used in Alpine environments, and has been calibrated from a large database of glaciers. The AAR method assumes that the accumulation area of an Alpine glacier is on average roughly 60-70% of the total glacier area, and thus allows an estimation of the ELA based on mapped extents (Gross and Kerschner, 1977; Maisch, 1992). By reporting ELA’s as a depression, or elevation relative to local LIA ELA’s, regional comparisons can be made that inherently include important effects of site-specific variables, such as valley morphology, precipitation distribution, and slope aspect. While such estimates are reasonably accurate for glaciers of simple geometry, the ELA depression is difficult to reconstruct for extensive or dendritic glaciers, and few ELA estimates have been reported for these systems.

Alpine Younger Dryas glaciers

Moraines associated with the Egesen stadial provide the most complete record of Younger Dryas ice extents in the central Alps. These moraines are characterized by a series of two, three, and in some cases four prominent, morphologically fresh ridges (Kerschner et al., 2000). Originally defined by a type locality in the Austrian Alps (Kinzl, 1929), the moraines were first related to a regional climatic event by Heuberger (1966, 1968), and have now been described throughout the French, Swiss, and Austrian Alps (Favilli et al., 2009; Federici et al., 2008; Herti and Kerschner, 2000; Kerschner et al., 2000; Hormes et al., 2008; Ivy-Ochs et al., 2006b,a, 2009; Kelly, 2003; Kelly et al., 2004; Kerschner et al., 2006; Maisch, 1982, 1992; Muller et al., 1980; Sailer et al., 1999; Scapozza et al., 2011; Schindelwig et al., 2012). At most locations outer Egesen moraines are sediment-rich, while the inner are composed of large blocks and boulders. The mark a distinct morphological
transition from relatively fresh LIA moraines up-valley to the slightly older, more subdued sediment-rich moraines of the Clavadel/Senders, and Daun stadia found at lower elevations (Ivy-Ochs et al., 2006b). Although this morphological and sedimentological contrast is in part caused by the different age of the moraines and post-depositional modification, it also likely reflects aspects of the valley’s glacial history prior to moraine formation.

Of the earlier stadia, at least the Gschnitz marked a true re-advance of glaciers through ice-free terrain (and not simply a glacier still-stand (Kerschner and Berktold, 1982; van Husen, 1977)), however, climate proxies indicate that ELA fluctuations were likely much smaller than those associated with the high-magnitude warming and cooling events at the Bølling/Allerød, Younger Dryas transition. Climatic conditions during the Bølling/Allerød interstadial (14.6-12.1 ka BP (Hajdas et al., 2004, 1993), see Figure B.2) were similar to those during the LIA, disrupting gradual Lateglacial warming and causing glaciers to retreat farther up-valley than the recent LIA extents (Ohlendorf, 1998), as the treeline rose above 1000 m a.s.l. for the first time since at least the mid-Würm glaciation (Furrer et al., 1987). Sedimentological evidence from Alpine lakes suggests that climatic fluctuations during the Bølling/Allerød likely resulted in temporary ice advances beyond LIA extents (Ivy-Ochs et al., 2008; Ohlendorf, 1998) before glaciers re-occupied tributary inner valleys during the cooler Younger Dryas interval. The relatively long (~2000 yr) Bølling/Allerød warm phase is therefore likely to have resulted in the most extensive glacier retreat and long-lasting interstadial since perhaps the mid-Würm glaciation. Subsequent initial Younger Dryas re-advances were similarly strong, and variable climatic conditions during this time lead to a number of minor retreat and advance phases, producing the characteristic nested sequence of moraines. Final retreat of glaciers following the Younger Dryas marked the onset of the Holocene, which was interrupted by a number of weaker advances, including the Kartell and previously mentioned late Holocene fluctuations (Ivy-Ochs et al., 2009; Kerschner et al., 2006; Maisch, 2000).
Local LGM and LIA

During the LGM, the Matter and Saas Valleys were occupied by the ∼1400 m thick Valais icefield, a major contributor to the Rhone Glacier (Jackli, 1970; Kelly, 2003; Penck and Brückner, 1909). LGM trimlines in the study region are generally located above 2600 m (Kelly et al., 2004), reflecting a western Alpine ELA depression on the order of 1200-1500 m (Keller and Krayss, 2008). While questions remain regarding the rate of post-LGM breakdown following ice retreat from the foreland, a compilation of pollen records and radiocarbon dates from lake and bog sites within western Switzerland (Wohlfarth et al., 1994) suggests the majority of the Rhone valley remained glaciated up until approximately 15.5 ka BP. The onset of sedimentation at an increasing number of sites after 14.5 ka BP suggests progressive deglaciation of principal inner valleys in the western Alps likely occurred immediately prior to, or synchronous with, the onset of the Bølling interstadial (Hajdas et al., 2004). Radiocarbon dates from the lowest humus layer in a peat bog at Zeneggen in the north of our study area (1510 m a.s.l, 900 m above the present-day Rhone valley) (Figure B.1), indicate that this region was ice-free by 15.0-13.9 ka BP (12.31±0.15 ka BP ^14C) (Welten, 1982, Figure B.2)). Pollen analysis from the sediment core indicates this sample corresponds to the Allerød phase, and the true depositional date is therefore likely to be at the younger end of this range (Welten, 1982). Recalculating ^10Be dates (see Section Sampling method and preparation) from glacially polished bedrock at an elevation of 2479 m in the southern Saas Valley (to the east of Saas Fee) (Kelly, 2003) yields a similar age of 15.2±0.2 ka BP (Figure B.1 and B.2), suggesting that thick glacial ice may have persisted in the valley until almost the onset of the Bølling interstadial. Both of these studies, however, represent minimum ages of deglaciation and may include significant lag-times.

The timing of the LIA is accurately known from historical records, and associated moraines have been mapped throughout the Matter and Saas Valleys (Maisch, 1992). These are common at elevations above 2800 m, although moraines from larger glaciers in the southern part of the area are found as low as 2000 m. Using the AAR method, Maisch (1992) determined LIA glacier ELA’s ranged from ∼2700 m in the north of the catchment to ∼3100 m in the vicinity of Monte Rosa.
B.2 Description of the study area

massif in the south (Figure B.1). For the purposes of this investigation, we assume a representative LIA ELA of 2700 m, which is based on an area-averaged mean of the mapped LIA glaciers in the valleys.

Local Lateglacial stadia

Early studies such as those in the Drance Valley (Burri, 1974), Valais (Bircher, 1982; Winstorfer, 1977), Val de Nendaz (Muller et al., 1980), Engadin (Maisch, 1981), Aletsch region (Welten, 1982), and greater Graubunden (Maisch, 1992) relied on comparisons of moraine morphology, pollen analyses, and radiocarbon dating of organic matter to identify and date morphological features representing Younger Dryas stadia. More recent investigations seeking to better constrain the timing and extent of Lateglacial advances in the Swiss Alps have benefitted significantly from developments in cosmogenic nucleide exposure dating, GIS technology, and digital terrain models. Such investigations at the Aletsch Glacier (Kelly et al., 2004), Grimsel Pass (Ivy-Ochs et al., 2006b), Julier Pass (Ivy-Ochs et al., 2006b), Mont Gele (Scapozza et al., 2011), and Belalp (Schindelwig et al., 2012) effectively utilized these modern techniques to accurately constrain Lateglacial ice extents within portions of the central Swiss Alps.

Winstorfer (1977) and Bircher (1982) investigated Late Pleistocene and Holocene climatic fluctuations in the Valais, and in particular the Matter and Saas Valleys. In a comprehensive sedimentological and morphological investigation, Winstorfer (1977) identified three possible Lateglacial stadia within the two valleys. Denoted Exterior, Interior, and Intermediate, the youngest of the mapped extents (Interior) was interpreted to correspond to the Daun stadial, with terminal moraines of both valley glaciers located close to the village of Stalden (Figure B.1). The most extensive stadial (Exterior) was inferred from ice-marginal sediments at Graechen, as well as glacial and fluvioglacial sediments in the Rhone valley itself (Figure B.1), suggesting this re-advance occurred while the valley glaciers extended beyond the confluence of the Matter and Saas Valleys. Bircher (1982) correlated moraines in the Saas Valley to five of the Alpine Lateglacial stadia based on moraine morphology and stratigraphy. He was able to date a peat deposit within the limits of a prominent Egesen moraine at Grosses Moos (at 1800 m a.s.l.) to 11.8-10.6 ka BP.
(GMO-1, 9.76±0.17 ka BP 14C) (Figure B.1 and B.2), constraining the timing of final retreat of Younger Dryas glaciers in the valley.

Recent investigations by Kelly et al. (2004) and Schoof (2010) used 10Be dating to resolve the depositional age of Younger Dryas moraines in the Aletsch glacier region to the north of our study area. Kelly et al. (2004) sampled four boulders on the crest of a Great Aletsch Glacier moraine (at ~2100 m, near Hohfluh), while Schoof (2010) presents twenty-one exposure dates for four individual moraines deposited by the Unnerbaech glacier at Belalp (Figure B.2). The moraine sampled by Kelly et al. (2004) displays classical Egesen characteristics, and a weighted mean of recalculated exposure ages suggests a depositional age of 12.6±0.1 ka BP, coincident with the earliest Younger Dryas cold event in the GRIP ice core (Figure B.2). Moraine morphology and stratigraphy indicate that all but the innermost moraine dated by Schoof (2010) represent Egesen stadia, and recalculated exposure ages suggest deposition of these moraines took place between 12.6±1.0 ka BP and 11.7±0.8 ka BP (samples VBA-23 and VBA-2), again consistent with the timing and duration of the Younger Dryas. These dates are supported by radiocarbon dates from bog sediments within the footprint of the Egesen stage Aletsch Glacier, which suggest that by 11.4±0.1 ka BP (12.3±0.15 ka BP 14C) ice had retreated at least 2 km up-valley and reduced in elevation by 120 m (Figure B.2) (Welten, 1982).

Paleo-ELA estimations for local Lateglacial stadia can be derived from a well-preserved sequence of moraines in Val de Nendaz, approximately 50 km to the west of our study area (Muller et al., 1980; Scapozza et al., 2011). This provides useful constraint on ELA depressions for Lateglacial stadia; which are summarized alongside age constraints for Alpine Lateglacial stadia in Table B.1.

B.3 Methods

B.3.1 Mapping

Identification of glacial features within the study region was undertaken using a combination of traditional field mapping, digital terrain model (DTM) analysis, and a review of existing literature. We limited our investigation to localities be-
### B.3 Methods

<table>
<thead>
<tr>
<th>Stage</th>
<th>Local sites</th>
<th>Age</th>
<th>ELA depression</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kartell stadial</td>
<td>Belalp,</td>
<td>10.8±1 ka BP</td>
<td>120 m</td>
<td>(Ivy-Ochs et al., 2006a; Nicolussi and Patzelt, 2001; Schindelwig et al., 2012)</td>
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<tr>
<td>Egesen stadial</td>
<td>Val de Nendaz, Belalp, Saas Valley, central Alpine Valleys</td>
<td>12.3±1.5-11.1±0.3 ka BP</td>
<td>I=300-150 m; II=70 m less than Egesen I; III=120 m less than Egesen I</td>
<td>(Bircher, 1982; Kerschner and Ivy-Ochs, 2008; Kerschner et al., 2000; Muller et al., 1980; Scapozza et al., 2011; Schindelwig, 2010)</td>
</tr>
<tr>
<td>Daun stadial</td>
<td>Val de Nendaz</td>
<td>Before Bolling</td>
<td>400-250 m</td>
<td>(Ivy-Ochs et al., 2006a; Maisch, 1982; Scapozza et al., 2011)</td>
</tr>
<tr>
<td>Clavadel/ Senders stadial</td>
<td>Val de Nendaz</td>
<td>Before Bolling</td>
<td>400-500 m</td>
<td>(Ivy-Ochs et al., 2006b; Maisch, 1982; Scapozza et al., 2011)</td>
</tr>
<tr>
<td>Gschnitz stadial</td>
<td>Val de Nendaz</td>
<td>(&gt;15.4±1.4 ka BP)</td>
<td>600-800 m</td>
<td>(Ivy-Ochs et al., 2008; Kerschner and Ivy-Ochs, 2008; Muller et al., 1980; Scapozza et al., 2011)</td>
</tr>
<tr>
<td>Last Glacial Maximum</td>
<td>European Alps</td>
<td>20.9±1.5ka BP</td>
<td>1000-1500 m</td>
<td>(Ivy-Ochs et al., 2006a; Keller and Krayss, 2008)</td>
</tr>
</tbody>
</table>

Table B.1: Approximate dates and ELA depressions for local Lateglacial stadia.
B. Reconstruction of Lateglacial stadia in the Matter and Saas Valleys, Canton Valais, Switzerland; implications for rock slope stability

tween the LIA (Maisch, 2000) and LGM (Kelly et al., 2004) extents (i.e. those interpreted to represent Lateglacial stadia), focusing on sedimentary and bedrock evidence within the main valleys that provide the best constraint on the maximum extent of Younger Dryas advances.

An initial desk study was undertaken combining previous mapping from Winkstorfer (1977), Bircher (1982), and the existing 1:25,000 Geological Atlas of Switzerland (Bearth, 1953, 1957, 1964, 1973, 1980). This was supplemented by a geographic information system (GIS) investigation, integrating recent 0.25 m resolution aerial photographs (Inc., 2011) and a digital slope-aspect map produced from high-resolution airborne LiDAR data (SwissTopo, 2009) within the Google Earth GIS platform. LiDAR data were collected using an aircraft-based laser unit to measure ground-surface elevations with an accuracy of greater than 0.5 m and horizontal resolution <2 m, providing good coverage of landforms even when partially obscured by vegetation. By converting elevation data into a slope-aspect map we were able to highlight contrasts in slope orientation, and in particular reversals in slope aspect associated with uphill-facing moraine ridges on the valley walls. LiDAR data are currently only available up to 2000 m elevation, sufficient to cover the lower 500 m of the main valley slopes and identify the most critical terminal and lower lateral moraines for the main valley glaciers. A compilation of moraines identified within the region is presented in Figure B.3.

Field investigations were undertaken during the summer of 2009/2010, focusing on morphological and sedimentological investigations of potential moraine locations, and identification of glacial features within the main valleys (such as polished or striated bedrock surfaces, postglacial sediments, and smoothed landforms). Qualitative comparisons of rock slope instability are based on observed talus distributions, topographic roughness (the irregularity of the rock slope surface), and rockfall activity within the valleys.

B.3.2 Sampling and age calculation

Site selection

We selected well-preserved moraines near the termini of a large dendritic paleo-glacier system for cosmogenic $^{10}$Be surface exposure age dating. Moraines relating
to the system were identified in both valleys during our mapping investigation (Figure B.3), and exposure dating allowed us to constrain their depositional age and correlate the stadia between valleys. The locations of sampled moraines are indicated in Figure B.3, and slope-aspect maps for each site are presented in Figure B.4.

The sampled Saas Valley moraine, located 1 km south of the village of Saas Balen at an elevation of 1650 m a.s.l., is delineated by a pair of ridges approximately 200 m long on the eastern flank of the valley (Figure B.4A). The ridges have a lateral separation of ~25 m, and are located below a rockfall deposit dominated by large angular (20-50 m$^3$) blocks. The upper moraine ridge is modified by a single
Figure B.4: Saas (A) and Matter Valley (B) moraine slope aspect map and sample locations. The colour model for the maps is represented as a lower hemispherical projection, where steep (90°) slopes are dark, and flat (0°) slopes are light. Slope orientation determines variation in hue; north-dipping slopes are blue, south-dipping are orange, and east- or west-dipping are represented by purple or green respectively.

lane road which bisects the surface, although it appears disturbance of boulders is minimal as the road was constructed largely on top of the existing deposit. Given its characteristic morphology, distinct sediment composition (including very large sub-rounded to rounded serpentinite and gneiss boulders), and proximity of this moraine to the terminus of the paleo-glacier, we select it as the best candidate in the Saas Valley, despite the previously mentioned sources of possible disturbance.

The selected Matter Valley moraine is located near the village of Randa at an elevation of 1750 m a.s.l. The moraine is a prominent ridge approximately 150 m long, with an uphill hollow, lying atop a small rock cliff. A blocky rockfall deposit, similar to that above the Saas Balen moraine, is located immediately upslope (Figure B.4B) which makes the assessment of provenance for boulders on
the crest complicated. We can, however gain some insight into the boulders origin by considering their depositional situation. They rest on, or are embedded within, pillars of silty to coarse blocky sediments (Figure B.5A), indicating boulder deposition has affected solifluction on the crest of the moraine, and they are likely to have been emplaced prior to stabilization of the moraine surface. We therefore assume that these boulders were deposited during, or shortly after, moraine formation and, due to their large size, have remained stable since.

Figure B.5: Photographs of dated Matter and Saas Valley boulders. These are (from left to right) the location of VM1a/b, boulder SB1, and boulder SB3.

**Boulder selection**

Two boulders, each measuring approximately 8 m$^3$, were sampled in the Saas Valley (samples SB1 and SB3, Figure B.5B and C), while one very large boulder ($\sim$60 m$^3$) in the Matter Valley was sampled twice (Figure B.5). Many candidate boulders at the Matter Valley moraine either showed signs of surface disturbance or low quartz content, and as a result we chose to sample only the single large gneiss boulder (samples VM1 and VM2). The two boulders on the Saas Valley moraine were located approximately 80 m apart in order to minimize the chance of sampling the same disturbance event (e.g., rockfall or road construction) (see Figure B.4A). SB1 was extracted from a gneiss boulder directly on the crest of the moraine (alongside the road), while SB3 was from another gneiss boulder slightly down-slope (in the trough between parallel moraines, see Section Moraine morphology). The single boulder selected on the crest of the Matter Valley moraine is approximately 2 m high and embedded in a mound of fine glacial sediment.
Sampling method and preparation

Samples were extracted from quartz veins located on top of each of the boulders using either a hammer and chisel or 30 mm diameter hole saw. Sample preparation followed the procedure described by Ivy-Ochs et al. (2006b). Samples were crushed using a disc mill to a target grain size of <0.8 mm (determined by sieving), and treated once with 32% hydrochloric acid and three times with 4% hydrofluoric acid, in order to purify the quartz component. Samples SB3 and VM1B each yielded approximately 50 g of quartz from 170 g and 200 g of crushed rock respectively, VM1a produced 25 g from 250 g of rock, and SB1 (a weathered, mica-rich sample weighing 170 g) yielded 8 g of quartz. Samples were processed in a single batch alongside a full-process blank containing the carrier solution. Relevant field and experimental data (corrected for blank $^{10}$Be concentration) are presented in Table B.2.

Measurement of $^{10}$Be content was undertaken at the ETH tandem accelerator mass spectrometer (AMS) facility in Zurich (normalized using the S2007N standard to a nominal $^{10}$Be/Be value of 28.1x10^{-12} (Kubik and Christi, 2010; Kubik et al., 2009; Muller et al., 2010)). Surface exposure ages were calculated assuming a $^{10}$Be half-life of 1.39±0.012-106 yr (Korschinek et al., 2010; Nishiizumi et al., 2007) and a cosmogenic $^{10}$Be production rate of 3.88±0.19 atoms $^{10}$Be g$^{-1}$ (determined for north-east North America, Balco et al. (2009), using the time-dependant model of Stone (2000)). Corrections to account for the sampled surface orientation and shielding of cosmic rays by topography were applied based on the method described in Dunne et al. (1999). No corrections for annual snow cover, palaeomagnetic intensity, or non-dipole effects were applied (the effect of palaeomagnetic variations at this latitude would have been less than 1% (Masarik et al., 2001)). A conservative erosion rate of 1.0 mm kyr$^{-1}$ was selected based on a comparison of boulder surface roughnesses resulting from variable quartz/feldspar weathering, and post-glacial weathering rates for biotite-rich granite derived by Andrè (2003). For consistency, all $^{10}$Be exposure ages quoted in this manuscript have been recalculated using the scheme described above, providing a standard reference for the calculated ages.
B.3 Methods

<table>
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<th>Lon.</th>
<th>Alt (m)</th>
<th>Thickness (cm)</th>
<th>Shielding correction factor</th>
<th>$^{10}\text{Be}$ ($\text{g}^{-1} \times 10^{-4}$)</th>
<th>Ave exposure age (yr)</th>
<th>Int. uncertainty (yr)</th>
<th>Ext. uncertainty (yr)</th>
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<tr>
<td>VM1B</td>
<td>46.096</td>
<td>7.789</td>
<td>1751</td>
<td>2.5</td>
<td>0.960</td>
<td>13.9±0.5</td>
<td>8670</td>
<td>320</td>
<td>530</td>
</tr>
<tr>
<td>SB1</td>
<td>46.145</td>
<td>7.931</td>
<td>1650</td>
<td>1.5</td>
<td>0.959</td>
<td>16.2±1.2</td>
<td>10820</td>
<td>820</td>
<td>970</td>
</tr>
<tr>
<td>SB3</td>
<td>46.144</td>
<td>7.930</td>
<td>1636</td>
<td>2.5</td>
<td>0.975</td>
<td>16.4±0.6</td>
<td>11180</td>
<td>430</td>
<td>690</td>
</tr>
</tbody>
</table>

Table B.2: Cosmogenic nuclide concentrations and calculated exposure ages.

B.3.3 Modeling

Concept of model

The numerical landscape evolution model ICE-CASCADE (Braun et al., 1999; Tomkin and Braun, 2002; Herman and Braun, 2008) uses the Shallow Ice Approximation (SIA Hutter, 1983) to solve the glacier mass balance equation. The model accounts for internal glacier deformation, basal sliding, snow and ice metamorphosis, surface accumulation and ablation, and melting at the glacier-bedrock interface. We do not enable any erosion or hillslope mass transfer within the model. Table B.3 lists inputs for our ICE-CASCADE model; all variables except the mean annual air temperature (MAAT) at sea-level remain constant. Progressive lowering of the model ELA (a function of the MAAT at sea-level and atmospheric lapse rate) is illustrated in Figure B.6, alongside model stages selected to reflect early Lateglacial stadia (Gschnitz, Clavadel, and Daun) and the LIA-equivalent stadial (see Section Correlation of modeled and mapped stadia).

Glacial ice development was initiated on a 200 m resolution present-day DEM, existing glaciers and valley sediment was not removed. The SIA favours well-developed valley glacier systems where the total ice thickness is greater than the local slope gradient, and model accuracy decreases at high slope-to-ice thickness ratios. Given the steep, high-relief terrain within the catchment, our model results for early stages (when ELA’s are relatively high and glaciers are thin) are less reliable. We use a sinusoidal function to drive ELA depression, allowing slow ice development in early stages to set initial conditions, which are then followed by an almost linear ELA decrease through levels appropriate for Lateglacial stadia –
B. Reconstruction of Lateglacial stadia in the Matter and Saas Valleys, Canton Valais, Switzerland; implications for rock slope stability

Figure B.6: Model ELA reduction and selected stages correlated to Lateglacial stadia. The dashed line indicates the area-weighted mean ELA of mapped LIA glaciers within the study region, while the shaded area represents their variance.

ensuring each simulated “re-advance” occurs under similar conditions. The ELA is reduced to our selected LGM-equivalent level over a period of 5000 years, then we hold the model inputs constant to ensure the model reaches equilibrium at the minimum MAAT. Glaciers at this stage of the model reach the edges of the model domain allowing free outflow of ice, so these results do not reflect true Alpine LGM conditions (see Section Modeled glacier advance for further discussion).

Input parameters

We use standard literature-derived input parameters for our model, and tune them to match the model ice coverage and ELA elevation of the mapped Younger Dryas ice extent. Selected model parameters are listed in Table B.3; additional details on the model physics and required inputs can be found in Braun et al. (1999); Herman and Braun (2008) and Tomkin and Braun (2002).
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
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<tbody>
<tr>
<td>DEM resolution</td>
<td>200 m</td>
</tr>
<tr>
<td>ICE resolution</td>
<td>300 m</td>
</tr>
<tr>
<td>Time step</td>
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<tr>
<td>Uplift rate</td>
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<tr>
<td>Max. ablation</td>
<td>$-20$ m yr$^{-1}$</td>
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<tr>
<td>Max. ice accumulation</td>
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</tr>
<tr>
<td>Atmospheric lapse rate</td>
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<tr>
<td>Max and min sea level temperatures</td>
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<tr>
<td>Period of surface oscillation</td>
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</tr>
<tr>
<td>Basal heat flux</td>
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<tr>
<td>Ice conductivity</td>
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<tr>
<td>Ice-flow constant</td>
<td>$6.8 \times 10^{-24}$ s$^{-1}$ Pa$^{-3}$</td>
</tr>
<tr>
<td>Sliding law constant</td>
<td>$5 \times 10^{-14}$ Pa$^{3}$ yr$^{-1}$ m$^{-2}$</td>
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<tr>
<td>Glen’s flow parameter</td>
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<tr>
<td>Exponent sliding law</td>
<td>3</td>
</tr>
</tbody>
</table>

Table B.3: ICE-CASCADE model input parameters.
B.4 Results

B.4.1 Age of mapped stadia

Results of laboratory analysis yielded consistent $^{10}$Be exposure ages for each moraine, although calculated ages for the two sites are distinctly different (Table B.2). Exposure ages from samples SB1 and SB3 correlate well within standard levels of uncertainty, and assuming the older sample SB3 reflects the timing of the Saas Valley moraine emplacement, we find that advance to the outer moraine ridge took place no later than 11.2±0.7 ka BP. We therefore correlate these moraines with Egesen age deposits identified elsewhere in the Alps (see Section Alpine Younger Dryas glaciers), representing a late Younger Dryas stadial (Figure B.2).

Samples VM1A and B from the Matter Valley moraine indicate an exposure age of 8.7±0.6 ka BP., significantly younger than both the Saas Valley moraine ages and the final Younger Dryas cooling event inferred from the GRIP ice core (Walker et al., 1999, Figure B.2). Although minor Alpine stadia have been correlated with 9.3 and 8.2 ka BP. cooling events (Nicolussi and Patzelt, 2001; Schindelwig et al., 2012), associated moraines are typically located near the LIA extent, and represent a much smaller advance than that mapped in this study. Results from our morphological and numerical studies (presented in the following sections) indicate that the dated moraines in each valley are likely to be chronologically equivalent, and when considered in combination with evidence of solifluction and rockfall deposits on the surface of the Matter Valley moraine, it is likely that exposure ages for this moraine reflect ongoing hillslope processes after deposition.

B.4.2 Moraine locations in the trunk valleys

Deposits interpreted to reflect terminal, or near-terminal moraines of the principal valley glaciers, suggest that the tongue of Egesen glaciers extended down to 1300 m elevation in the Matter Valley, and 1650 m in the Saas Valley (Figure B.3). These locations are 10 to 15 km beyond the maximum extent of the lowest LIA glacier termini (those of the Gorner and Allalin Glaciers). Furthermore, lateral moraine deposits located in both valleys lie between 300 m and 500 m above the present-day valley floors, and along with trimlines and moraines mapped in
tributary valleys, delineate an extensive dendritic Egesen glacier system within the region. The predicted extent of the glaciers, as well as the location of potential moraines identified in previous investigations, is illustrated in Figure B.3. Our interpreted Matter Valley terminal moraine is illustrated in Figure B.7, and highlights the benefit of high-resolution LiDAR data used in this investigation. Although largely overrun with talus and alluvial sediment, subtle features downstream of the proposed terminal moraine location are textured differently to the talus, or gently slope away from the valley axis. Although it is difficult to interpret sedimentary evidence in the field at this location, we suggest that the subtle morphologies represent at least two distinct ridges or outwash terraces associated with a glacier terminus within the valley.

Figure B.7: Presumed Matter Valley terminal moraine or glacial outwash deposit location. Arrows indicate traces of outward-dipping ridges. Smoother, “lumpy” regions are inferred to reflect glacial sediments, while talus associated with rockfalls is generally rougher and more linear in appearance.

B.4.3 Moraine morphology

Moraines of the Egesen stadial in each valley are characterised by a distinct rounded outer ridge (approximately 5 m wide), with a smaller (~3 m wide) inner
B. Reconstruction of Lateglacial stadia in the Matter and Saas Valleys, Canton Valais, Switzerland; implications for rock slope stability

ridge or flattened terrace ~10 m downslope. Large, blocky (10-100 m$^3$) rockfall deposits, possibly remnant from ongoing slope collapse during LGM ice retreat, are commonly found beyond the outer ridge. The largest boulders within the moraines are typically orthogneiss or paragneiss material, reflecting the dominant lithology of the region, and in the case of the sampled Matter Valley moraine, are elevated from the moraine ridge by a “pillar” of fine sediment likely resulting from post-depositional solifluction.

Figure B.4 shows slope-aspect maps for the sampled Matter and Saas Valley moraines derived from aerial LiDAR data. These maps illustrate the linear outer (a) and inner (b) moraine ridges on the eastern side of the Saas Valley, close to the termination of the interpreted valley glacier. The steep slope below the main crest of the sampled Matter Valley moraine is likely to have prevented preservation of the inner ridge at this location, although it is observed at corresponding locations farther up-valley. Clearly visible to the north of the Matter Valley site is a distinct ridge perpendicular to the slope and associated with a deep-seated, post-glacial rock slope instability that has displaced talus on the slope. It is likely this pre-existing ridge enabled the preservation of our dated moraine on the otherwise steep valley wall.

B.4.4 Modeled glacier advance

We present a summary of modeled ice extents for 100 m intervals of ELA depression in Figure B.6, providing an indication of the progressive development of valley glaciers in response to ELA reduction. Although Lateglacial advance and retreat cycles were significantly more complicated than the simplified cooling scenario implemented in our models, the progression provides a useful indication of the effect of catchment geometry on glacier development, aiding identification of locations where glacial fluctuations may have been more sensitive to ELA variation. Model results predict a non-linear increase in down-valley ice progression until the ELA reaches approximately 2450 m, below which propagation of the glacier tongue settles at a consistent rate of approximately 6 km per 100 m ELA decrease (Figure B.8). The slow development of valley glaciers in response to initial ELA reduction (toward 2450 m) implies that a significant climatic driver is required to progress
glaciers into the main trunk valleys, and ice is therefore likely to be restricted to the upper half of each valley during typical interstadial conditions.

Figure B.8: Modeled glacier extents for 100 m ELA intervals. We present the full range of model results, although it is likely that model results do not reflect the true extent of lateglacial ice within the valleys for stages where the glacier tongue extends beyond Visp (∼2100 m ELA, see Section Modeled glacier advance for further discussion).

Glaciers from the Matter and Saas Valleys converge at Stalden as the ELA decreases to 2240 m, and extend out to the Rhone valley once the ELA reaches 2150 m (an equivalent ELA depression of ∼550 m, see Figure B.6 and Figure B.8). Although we model ice extents for ELA’s down to 1300 m, the spatial extent of our model is limited, and as such, we cannot account for ice sourced from the headwaters of the Rhone glacier east of Visp. This was one of the two largest glaciers in the Swiss Alps during the LGM, and it is likely that ice persisted in the Rhone valley for a significant period of time following retreat from the Alpine...
B. Reconstruction of Lateglacial stadia in the Matter and Saas Valleys, Canton Valais, Switzerland; implications for rock slope stability

foreland, possibly until the Bølling warm period (see Section Local LGM and LIA). It is therefore likely that the Rhone glacier affected the dynamics of the Matter and Saas glaciers until they retreated south from Visp. This interaction would tend to reduce ice flow velocities within the two valleys, increasing ice accumulation and therefore the elevation of ice in the valleys south of Visp. Modeled ice extents for ELA’s less than 2150 m should therefore be considered as minimums, and, as illustrated by a comparison of the modeled results to mapped LGM trimlines (Kelly et al., 2004, Figure B.8), the true LGM elevation within the region is likely to be between 200 m and 300 m higher than we model.

B.5 Discussion

B.5.1 Correlation of modeled and mapped stadia

Correlations of our model results with mapped Alpine Lateglacial stadia are presented in Figure B.8. As expected, the model does not perform well for small, steep LIA glaciers, and modeled ice extents during warmer stages correlate poorly to mapped glaciers. The best-fit model stage for LIA glaciers reflects an ELA that is 134 m above the mean regional LIA ELA (see Section Local LGM and LIA), within the upper 1/3 of ELA’s derived for LIA glaciers in the region, though strongly biased toward the northern glaciers and not representative of the valley as a whole. We instead use the ELA calculated from mapped glaciers for stadial correlations (see Section Local LGM and LIA).

Despite significant simplifications and limited spatial resolution (see Section Concept of model), the modeled Egesen Stadial ice extents indicated in Figure B.9 are consistent with our mapped moraines (although the sampled Matter Valley moraine is perhaps too far behind the glacier terminus to definitively link it to this advance), and the model ELA depression for this stadial correlates reasonably well with field observations (see Figure B.6, Table B.1). A comparison of the modeled ELA for a stage matching the inner of two prominent moraines to the north of Saas Fee (Figure B.9), indicates that the moraine is likely related to a late Egesen II (or III) re-advance, although this is poorly constrained throughout the Alps.
B.5 Discussion

Figure B.9: Model stages and mapped moraines used to correlate results to Lateglacial stadia. Apparent is the poor fit for mapped and modeled LIA glaciers (see Section Correlation of modeled and mapped stadia), though the correlation with mapped Younger Dryas ice extents is very good. In particular, two moraines north of Saas Fee provide confidence on our interpretation of Egesen I and III stadial extents.

Reduction of the model ELA from Egesen-equivalent conditions toward that of the regional LGM results in the progressive development of ice in the valleys, allowing us to correlate modeled ice extents with various mapped (though undated) moraines and Lateglacial ELA depression estimates from nearby valleys (see Sections Local Lateglacial stadia and Mapping). Our results suggest that glaciers during the Gschnitz stadial are likely to have extended out to the Rhone valley, while the Clavadel stadial would terminate just south of Visp, and Daun-equivalent glaciers are likely to have terminated immediately north of Stalden (Figure B.9). Our model results are remarkably similar to the interpreted stadia of Winstorfer (1977), who also suggested that the most extensive Lateglacial stadial (our
Gschnitz) extended out into the Rhone valley.

The confluence of modeled Daun-equivalent glaciers near Stalden is similar to the mapped “Intermediate” stage glaciers of Winstorfer (1977), however, sedimentary evidence at this location is difficult to interpret. Winstorfer (1977) suggests that the Saas glacier is likely to have extended farther down-valley than the Matter glacier at this time-based largely on the presence of significantly more glacial sediments in the lower Saas Valley. However, outcrops at the confluence commonly show signs of having been deposited within a fluvial system, and may instead be remnant kame terraces, or reflect proglacial sediments impounded in the Saas Valley by the Matter glacier. While optimizing the model, we observed that increasing the annual precipitation rate tended to force the Matter glacier farther down-valley than the Saas, causing the glacier to reach the confluence earlier.

Our maximum modeled ice elevation in the upper Saas Valley is approximately 50 m below the bedrock surfaces dated by Kelly (2003) opposite Saas Fee (see Section Local LGM and LIA), and as discussed in the previous section (Section Modeled glacier advance), we suggest this region of the slope is therefore likely to have been glaciated only as long as the main Rhone glacier controlled ice flow out of the valleys (approximately the time of our correlated Gschnitz stadial). Ice on this slope reaches a maximum elevation once the model ELA drops below 2320 m (i.e. a depression of 380 m, Figure B.8), although our glaciers do not reach the Rhone Valley until the ELA drops to 2150 m. This introduces uncertainties to the interpretation of these dates, and while we consider it unlikely the slope was glaciated during the Clavadel advance (with a model ELA depression of 535 m), or Daun stadial (model ELA depression of 440 m), our model results do not allow us to exclude these possibilities. We suggest recalculated bedrock exposure ages of 15.2±0.2 ka BP (Kelly, 2003) therefore reflect cumulative exposure since the Gschnitz stadial, and assuming limited surface erosion, provide a minimum age for the stadial.

B.5.2 Lateglacial reconstruction

The results of this study provide important insights into the Lateglacial history of the Matter and Saas Valleys, in terms of both fluctuating valley glaciers, and
B.5 Discussion

deglaciation of the bedrock slopes. Following the LGM at approximately 21 ka BP, a gradually warming climate drove ice retreat from the Alpine foreland back toward the frontal Alpine ranges. The Rhone Glacier was confined to inner Alpine valleys by 16.1 ka BP (Section Alpine Lateglacial), though retreat from the foreland most likely occurred prior to this. The Rhone Valley then remained glaciated up until approximately 15.5 ka BP (Section Local LGM and LIA). Downwasting of ice in the Matter and Saas Valleys from the LGM until this time is unlikely to have exceeded 200 m. The Gschnitz stadial marks a temporary reversal of glacier retreat as ice re-advanced into the Rhone valley (Figure B.10), or possibly remained connected to a downwasting Rhone glacier. Significant downwasting throughout the valleys is likely to have followed the Gschnitz stadial, as glaciers retreated out of the Rhone valley and up toward Stalden, before potentially re-advancing toward Visp as the ELA again dropped to 535 m below the LIA level. This latter assumed Clavadel equivalent advance is likely to have terminated close to Visp (Figure B.10), and was followed by ice retreat back into the Matter and Saas Valleys. A subsequent re-advance towards Stalden as the ELA depression approached 440 m then led to the formation of the Daun stadial. Data to constrain the degree of glacial retreat prior to each of these stadia are currently limited, however, if we assume the Gschnitz advance occurred at the beginning of the Oldest Dryas (~16.6 ka BP), we can bracket a 2000 year pre-Bølling interval of the GRIP δ¹⁸O curve for the Gschnitz, Clavadel, and Daun stadia. The periodicity and magnitude of fluctuations in this section of the GRIP curve are relatively consistent (Figure B.2), as is the modeled sensitivity of our valley glaciers within the region (i.e. ±10 km of Stalden, Figure B.8). Our model suggests that fluctuations of the glacier tongue position are likely to have been relatively rapid within this time range, however, we propose it is unlikely that rock slopes in this lower valley region were exposed for much more than 500 years prior to each re-advance.

Initial deglaciation of the upper Matter and Saas Valleys is likely to have taken place following the Daun stadial, coincident with the onset of the Bølling warm period. Again we have few data to constrain the rate or degree of retreat during this period, although based on the GRIP δ¹⁸O record and studies of Bolling/Allerød glacier extents elsewhere in the Alps, it is likely that this was a significant retreat approaching the recent LIA extents in the valleys (Section Alpine Younger Dryas
B. Reconstruction of Lateglacial stadia in the Matter and Saas Valleys, Canton Valais, Switzerland; implications for rock slope stability

![Figure B.10: Model reconstructions of Lateglacial stadia (ice contours represent 100 m intervals).](image)

Our SB3 exposure age and Bircher’s (1982) $^{14}$C date at Grosses Moos provide some indication of potential interstadial glacier retreat rates within the Saas Valley, as they bracket 5 km of valley glacier retreat at the end of the Egesen stadial. These dates are indistinguishable within the range of error; however, the considerable ($\sim$1000 year) error means we can constrain the retreat to a minimum estimated mean rate of 5 m yr$^{-1}$. Volume-based calculations described by Raper and Braithwaite (2009) provide an additional first-order estimate of the likely response of the glaciers to this warming, and suggest ablation of ice in the valleys in response to a Clavadel to LIA-equivalent ELA rise is likely to require no more than a few hundred years. We therefore assume glacier retreat kept pace with the $\sim$350 yr period of rapid warming at the beginning of the Bølling interval (see Figure B.2). Given an approximately 2000 yr interval for the Bølling/Allerød, bedrock valley walls are therefore likely to have been exposed for at least 3-4 times as long prior to the Egesen stadia than during previous interstadia. This exposure will have provided significantly more time for weathering and fracture development on the rock walls, degrading surface bedrock, and promoting rockfall activity and displacement of deep-seated slope instabilities.
The climatic cooling at the onset of the Younger Dryas (approximately 12.7 ka BP) initiated re-advance of glaciers back into the trunk valleys during the Egesen stadial (Figure B.10). We suggest these advancing glaciers were able to quarry and clean loose rock, trigger rockfall activity, and undermine the toe of deep-seated instabilities much more efficiently than those during the higher-frequency advance/retreat cycles of the earlier Lateglacial stadia. These former glacial advances progressed through a younger, transitional landscape with a greater proportion of remaining post-glacial sediment and less-developed rock slope instabilities. As a result, Egesen moraines in the valleys can contain a greater proportion of rockfall-sourced material than the finer till characteristic of earlier moraines. Rock slopes within the footprint of the Egesen stadial can also be expected to appear “cleaner”, with more smoothed glacial surfaces surviving than those down-valley that did not share a similar history. Instabilities developed on rock slopes farther down-valley, following the Daun stadial, have not been affected by subsequent glaciation, and as a result much of the unstable bedrock that developed post-glacially remains in situ, or has contributed to the development of large talus fans within this region.

When compared to both the (~100 kyr) duration of the Würm glaciation and recent (7-8 kyr) Lateglacial period, the duration of the Younger Dryas advance and Egesen stadial is relatively short (~1000 yr). However, our observations suggest that glaciation of the Matter and Saas catchments during the Younger Dryas had a remarkable impact on present-day valley geomorphology. Although the residence time and number of glacial advance/retreat cycles clearly influences the ability of glaciers to erode and transport sediment, this pronounced effect following the Bølling/Allerød warm interval indicates that the interplay of glacial and bedrock fracture processes had a significant effect on glacial erosion and post-glacial rock slope development.

B.6 Conclusions

We combined \(^{10}\)Be surface-exposure dating with geomorphic mapping and numerical modeling of glacial ice extents to unravel the glacial history of the Matter and Saas Valleys in southern Switzerland. A key component of our study was the
identification and mapping of an extensive moraine sequence extending between 10 km and 15 km beyond the maximum extent of LIA glaciers in the trunk valleys. Surface exposure ages from the crest of a gneiss boulder located between nested moraines near the terminus of the mapped Saas Valley glacier indicate deposition occurred prior to 10.8±1.0 ka BP. In combination with observations of moraine morphology, this measurement allowed us to correlate the moraine sequence to the maximum Egesen stadial during the Younger Dryas cold interval. A calibrated numerical model of glacial ice extents (created using the code ICE-CASCADE) supports this correlation, as the stadial mapped in the two valleys is accurately reproduced once the model ELA depression reduces to 270 m below the mean LIA ELA determined from glaciers in the area. This value is in agreement with calculated ELA depressions for smaller Egesen stadia mapped elsewhere in this region of the Alps. Reducing the modeled ELA toward that of the LGM allowed us to reproduce the Daun, Clavadel, and Gschnitz Lateglacial stadia, which were also correlated with mapped moraines in the valleys. Our results indicate that these earlier stadia were significantly more extensive than the Egesen, and ice terminus locations may have been more sensitive to climatic fluctuations than those of Egesen or smaller glaciers in the valleys. The distribution of slope instabilities and glacial landforms within our study region indicates that Younger Dryas glaciers were more effective erosional agents than those of earlier Lateglacial stadia. We suggest this is a result of increased bedrock fracturing and rock slope weathering during the long Bølling/Allerød warm interval that preceded the Egesen advance.
Bibliography


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Acknowledgements

It is impossible to thank explicitly all the people to whom I am grateful for having accompanied and supported me during these years in Zürich. All those with whom I have discussed geology (in front of a computer or a beer), but also just chatted, traveled, hiked, sang, eaten rösti or spätzli, drunk, played pétanque, skied, had a tea, gone to conferences, swam, danced (!?), gone on holidays, played football, watched football, played the ukulele, laughed, seen a movie, worked, said nonsense, gone to the field, had a coffee, been at the Safari bar, uncorked a bottle, been at the river, cheered and “charlesed”, climbed, gone to the lake, played the guitar, ..., have contributed to make my life in Zürich particularly enjoyable. Many, many thanks to all!!

There are, however, some people that I want to thank in particular. The first one is Fred who gambled on me in spite of an awful interview. The vast majority of what I learned in these years comes from him and I did learn a lot. Hidden in his cryptic way of saying things without saying them, I found precious advices that led me through the difficulties of becoming a scientist. In any occasion, he has listened to me, being open and helpful. I think of him not only as a mentor, but especially as a good friend.

I want to thank Sean, who has given me the opportunity of making this experience but also thought me that being sharp and direct is often a fruitful approach. Sean has been to me like one of those uphill climbs in which you thought you have almost reached the top but when you roll the bend you still see a long, long way to go. It does not take long to understand that the goal is not to reach the top but rather to walk the climb and enjoy it.

Particular thanks goes to my office mates. Stefan, who reminds me of Switzerland, not only because of his nationality, but also because of his surprising pleasantrity. Rebecca, the office Queen. She made it to share the office with four man, a proof of her strength and determination she also shows in science. Teo, probably the most genuine person I have ever met. His authenticity and naturalness really makes him special. And, finally, Matt: in spite of the many differences that characterize us, you have been one of my closest friends since the beginning. I have wonderful memories of everything we did together. Getting to know you has
been a real pleasure. I wish that the end of this phd does not mean the end of the friendship I have with all of you.

Many thanks to Jean-Daniel and Pierre, who have been at the same time colleagues and friends. Working with them is way more a pleasure than a duty.

A special thanks to Achille. Having spent almost nine years together, I will never get used to not have him around. Being with him is one of the greatest pleasures that one can ever enjoy.

I am grateful to my family who has never stopped believing in me and supporting me in any possible way. I am aware that they silently suffered my absence. Whatever the future, I hope I will be able to fill the gap better than I have done during these years.

Finally, I owe infinite gratitude to Laura who has been waiting for me way too much time. I could not write here anything more effective than show her my love every day. I cannot wait to join her and start doing it.
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  · Development and application of a large-scale numerical model to simulate fluvial and glacial erosion.
  · Numerical reconstruction of past topographic configurations.
  · Morphometric analysis of the present-day topography.

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· Computation of the topographic error and surface displacement rates.
· Geological characterization and interpretation of the surface displacements.

State University of Milano, Earth Science Department “Ardito Desio”, Italy
Advisors: Bruno Crippa and Roberto Sabadini
Project: Influence of the atmospheric delay in the analysis of surface displacements using radar differential interferometric techniques (DInSAR).
· Processing satellite data.
· Quantification of the atmospheric delay on radar wave propagation rates.
· Implications for geological application of DInSAR techniques.

PUBLICATIONS

Published:

In revision:
· Leith K., Moore J. R., Ivy-Ochs S., **Sternai P.**, Loew S., Reconstruction of Lateglacial stadia in the Matter and Saas Valleys, Canton Valais, Switzerland; implications for rock slope stability. Submitted to *Quaternary Geochronology*.

In preparation:
· **Sternai P.**, Herman F., Valla P. G., Champagnac J.-D., Geomorphic control and headward propagation of glacial erosion in the Rhône valley.
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· Assistent in BSc course: Geological Mapping, fall sem. 2008

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· Socrates Erasmus, awarded to undergraduate students at the State University of Milano, 9.2007-2.2008

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· Geology/geomorphology/geophysics
· Surface process and geodynamic modeling
· Finite Difference and Finite Elements Methods
· Inverse methods
· 3D visualization
· Programming: bash, Fortran, Matlab, GMT
· Computer skills: Unix, Linux, OS-X, Window, MS Office, Adobe CS, LaTex, Matlab, ArcGis and many others
· Lidar differential GPS surveying
· Languages: Italian (native), English (fluent), French (fluent)

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