Computers & Geosciences 94 (2016) 77–85

Contents lists available at ScienceDirect

Computers & Geosciences

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

Research paper

GlaRe, a GIS tool to reconstruct the 3D surface of palaeoglaciers

Ramón Pellitero [a,*], Brice R. Rea [a], Matteo Spagnolo [a], Jostein Bakke [b], Susan Ivy-Ochs [c], Craig R. Frew [a], Philip Hughes [d], Adriano Ribolini [e], Sven Lukas [f], Hans Renssen [g]

[a] Department of Geography and Environment, University of Aberdeen, St Mary's, Elphinstone Road, AB24 3UF Aberdeen, United Kingdom
[b] Department of Earth Science, University of Bergen, P.O.Box 7800 5020 Bergen, Norway
[c] Insitut für Teilchenphysik, ETH-Höggerberg, Otto-Stern-Weg 5, 8093 Zürich, Switzerland
[d] Geography, School of Environment, Education and Development, University of Manchester, Oxford Road, Manchester M13 9PL, United Kingdom
[e] Dipartimento Scienze della Terra - Università di Pisa - Via S, Maria 53, 56126, Italy
[f] Queen Mary University of London, School of Geography, Mile End Road, London, E1 4NS, United Kingdom
[g] Department of Earth Sciences, Faculty of Earth and Life Sciences, VU University, Amsterdam, Netherlands

ARTICLE INFO

Article history:
Received 23 November 2015
Received in revised form
18 May 2016
Accepted 6 June 2016
Available online 11 June 2016

Keywords:
Glacier reconstruction
Flowline equilibrium profile
GIS
Python

ABSTRACT

Glacier reconstructions are widely used in palaeoclimatic studies and this paper presents a new semi-automated method for generating glacier reconstructions: GlaRe, is a toolbox coded in Python and operating in ArcGIS. This toolbox provides tools to generate the ice thickness from the bed topography along a palaeoglacier flowline applying the standard flow law for ice, and generates the 3D surface of the palaeoglacier using multiple interpolation methods. The toolbox performance has been evaluated using two extant glaciers, an icefield and a cirque/valley glacier from which the subglacial topography is known, using the basic reconstruction routine in GlaRe. Results in terms of ice surface, ice extent and equilibrium line altitude show excellent agreement that confirms the robustness of this procedure in the reconstruction of palaeoglaciers from glacial landforms such as frontal moraines.

1. Introduction

The existence and dimensions of terrestrially terminating glaciers is controlled, to a first order, by climate, through variations in temperature and precipitation (Ohmura et al., 1992). The Equilibrium Line Altitude (ELA) of a glacier is the key glacier surface location where empirical relationships relating these two climatic parameters have been determined (Ohmura et al., 1992; Braithwaite, 2008). The ELA is the elevation on the glacier where, at the end of the ablation season, the net mass balance is zero (i.e. snow and ice melted equals snow and ice accumulated within one year). ELAs can be calculated on present-day glaciers by making surface mass balance measurements (measured or geodetic) and can also be determined for palaeoglaciers using various techniques, most of which require knowledge of some component of the glacier geometry (Pellitero et al., 2015). Once a palaeoglacier ELA has been established, either the temperature or precipitation relating to it can be determined, providing quantitative palaeoclimate information for that location (Hughes and Braithwaite, 2008). Given the importance of the ELA for palaeoclimatology especially in high latitudes and altitudes, a rigorous reconstruction of the 3D geometry of the palaeoglacier is essential (e.g. Carr et al., 2010).

Palaeoglacier surface and volume reconstruction methods rely on morphological evidence of the former glacier geometry (e.g. terminal and lateral moraines, lateral meltwater channels, trimlines, kame terraces and ice contact deltas), to either initiate (iterative) or constrain (dynamic) the model (Federici et al., 2008; Lukas, 2006; Pellitero, 2013; Rea and Evans 2007). Ideally, there would be a wealth of landform evidence available for the reconstruction but in reality, most landforms are missing or fragmentary, especially in the accumulation zone, and often become increasingly degraded with age (e.g. Dawson, 1979). The best practice is therefore to use the available landform evidence in combination with numerically derived reconstructions. The numerical approach is rooted in the constitutive equations for glacier motion (Nye, 1952a, b) and creates an equilibrium glacier profile over the known, former, subglacial bed. This approach makes three assumptions:

1. the present-day topography is the same as the palaeoglacier basal topography. Evidence of considerable post-glacial

* Corresponding author.
E-mail addresses: ramon.pellitero@gmail.com (R. Pellitero), b.rea@abdn.ac.uk (B.R. Rea), m.spagnolo@abdn.ac.uk (M. Spagnolo), Jostein.Bakke@uib.no (J. Bakke), ivy@phys.ethz.ch (S. Ivy-Ochs), craigrfrew@gmail.com (C.R. Frew), Philip.Hughes@manchester.ac.uk (P. Hughes), ribolini@dst.unipi.it (A. Ribolini), s.lukas@qmul.ac.uk (S. Lukas), h.renssen@vu.nl (H. Renssen).

http://dx.doi.org/10.1016/j.cageo.2016.06.008
0098-3004/

geomorphological activity by proglacial, periglacial, paraglacial, fluvial and/or mass movement processes (e.g. infilled lakes or large mass movements) should be taken into consideration, and where possible, the present-day topography corrected.
2. the reconstructed glacier was in equilibrium with climate.
3. the palaeoglacier was land terminating. Calving impacts on the mass balance via geometrical and mass flux changes.

This paper presents a GIS tool that semi-automatically reconstructs the 3D geometry for palaeoglaciers given the bed topography. The tool utilises a numerical approach and can work using a minimum of morphological evidence (i.e. the position of the palaeoglacier front, a lateral moraine or a trimline). The numerical approach is based on an iterative solution to the perfect plasticity assumption for ice rheology, explained in Benn and Hulton (2010). The tool can be run in *ArcGIS 10.1 (ArcInfo license)* and later updates and the toolset is written in Python code.

2. Perfect plasticity rheology

The model implemented in GlaRe produces an equilibrium profile of a glacier in two dimensions (i.e. along the central flowline). The model takes no account of basal sliding, and it assumes that ice has a perfect plasticity rheology (Paterson, 1994, p. 240). It is based on the Shilling and Hollin (1981) equation:

$$h_{i+1} = h_i + \frac{\tau_{av}}{F_i \rho g}\frac{\Delta x}{H_i} \tag{1}$$

where, h is ice surface elevation, τ_{av} is basal shear stress (in Pa), F is a shape factor, ρ is ice density ($\sim$900 kg m^{-3}), g is the acceleration due to gravity (9.81 ms^{-2}), Δx is step length (in metres), H is ice thickness (in metres), and i refers to the iteration (step) number. This is a derivation from the well-known Nye (1952a) formula for the calculation of shear stress at the base of a glacier

$$\tau = \rho g H \sin\alpha \tag{2}$$

where, τ is the basal shear stress, and α is the ice surface slope.

Eq. (1) does not have a solution at the snout of the glacier, because τ_i and τ_{av} are equal to 0. Van der Veen (2013) solved this shortcoming by evaluating the ice thickness and the shear stress at the midpoint along iterative steps. The result is Eq. (3), which can be solved as a quadratic equation. The complete explanation and development of these formulae can be found in Benn and Hulton (2010) and Van der Veen (2013).

$$h_{i+1}^2 - h_{i+1}(b_i + b_{i+1}) + h_i(b_{i+1} - H_i) - \frac{2\Delta x \bar{\tau}_{av}}{Fg} = 0 \tag{3}$$

where, b is the bed elevation and the overbar indicates that the yield stress is the average for the interval.

In this paper we present a GIS tool that utilises this numerical approach to reconstruct the geometry of former, land-terminating glaciers, provided that the position of their frontal moraine or ice margin is known. It is suited to the reconstruction of cirque and valley glaciers and it can also be successfully used for plateau-fed glaciers and small ice caps/fields. The GIS tool allows users to define three input parameters – the basal shear stress (τ), the shape factor (F) and the interpolation procedure. These inputs are discussed further below.

2.1. *Basal shear stress τ*

The model requires the glacier basal shear stress as a primary input because this parameter exerts a first-order control on the output glacier 3D surface. Field and experimental data indicate that τ should lie in the $\sim$50–150 kPa range (Nye, 1952b) for a valley glacier (Paterson, 1970) and up to 190 kPa for a cirque glacier (Weertman, 1971). The ideal situation is to initially reconstruct the ice surface using a standard value of 100 kPa and then tune τ to fit the reconstructed 3D glacier surface to the geomorphological constraints on vertical ice thicknesses (e.g. lateral moraines and/or trimlines) (Benn and Hulton, 2010; Shilling and Hollin, 1981). However, these landforms are seldom present or difficult to identify and generally not easy to link to a specific frontal moraine. In the absence of any constraining geomorphology for palaeoglacier thickness, the standard reference value of 100 kPa (Paterson, 1994; Rea and Evans, 2007) is recommended and is the default for the tool. Shear stress may not be uniform along the length of a glacier because of bed changes along the flowline (e.g. a change in bedrock lithology/roughness, sediment cover across the valley floor or the decrease of shear stress near the ice divides of plateau icefields/ice caps, see Section 3). Therefore, experienced users are also given the option to manually alter the shear stress at any point along a flowline.

2.2. *Shape factor (F factor)*

The Nye (1952a) equation assumes that all the glacier driving stress is supported by the basal shear stress. This is unrealistic for valley glaciers and other topographically constrained glaciers (Benn and Hulton, 2010; Nye, 1952b; Shilling and Hollin, 1981). In these cases, significant resistance to flow can also be provided from lateral-drag, which can be incorporated into the formula using a shape friction factor, the *F* factor.

The concept of *F* factor was first introduced by Nye (1952b), who suggested it to be a function of the cross-sectional area and perimeter length (equivalent to the wetted perimeter of a river). The *F* factor was further discussed by Nye (1965), who reduced it to a function of the glacier width and thickness, for simple cross-sectional geometries:

$$W = \frac{w}{H} \tag{4}$$

where, for any cross-section, W is a shape-factor indicator, w is half the width of the glacier and H is the centre line ice thickness. W values are converted to the relevant F factor value using conversion tables (Li et al., 2012; Nye, 1965; Paterson, 1994, p. 269). However, W is an approximation of the true shape factor, as the calculation is based on simplified valley cross-sections (i.e. parabolic, semi-ellipse or rectangular). The original F factor is both a more sophisticated and geometrically correct approach and is the preferred method. It utilises the cross-sectional area and perimeter (Shilling and Hollin, 1981). If we extend Eq. (2) to the glacier perimeter at a glacier cross-section we find that

$$\tau = \rho g A \sin\alpha \tag{5}$$

where, A is the cross-section area. Assuming the driving stress is equal to the basal shear stress at the centre line, then the F factor is calculated by the following equation,

$$F = \frac{A}{Hp} \tag{6}$$

where, H is the ice thickness at a point and p is the length of the cross section ice-bed contact.

This is the approach taken by Benn and Hulton (2010) and implemented in the reconstruction tool here. The F factor, as defined in (6), should be equal to 1 for icecaps and icefields, where there are plateau source areas or poorly defined valley heads, as the driving stress here is entirely supported by basal shear stress (i.e. they are not topographically constrained). However, its value