Quaternary Science Reviews 80 (2013) 169–194

Contents lists available at ScienceDirect

Quaternary Science Reviews

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

New perspectives on paleoglaciology

J.L. Fastook [a], T.J. Hughes [b,*]

[a] Department of Computer Sciences, Climate Change Institute, University of Maine, Orono, ME, USA
[b] School of Earth and Climate Sciences, Climate Change Institute, University of Maine, 404 North Sixth Street, Fort Pierre, SD, USA

ARTICLE INFO

Article history:
Received 16 May 2013
Received in revised form
25 August 2013
Accepted 29 August 2013
Available online 8 October 2013

Keywords:
Quaternary Ice Age
Glaciation cycles
Sea ice
Ice shelves
Ice sheets
Glacial geology
Marine geology
Arctic Ocean
White Hole
Megafloods

ABSTRACT

Paleoglaciology deals with glaciation cycles of the Quaternary Ice Age. It combines the dynamics of present-day ice sheets deduced from glaciology with the history of former ice sheets deduced from glacial geology. Cosmogenic dating now makes a detailed chronology of a cycle possible. Radiocarbon dating has provided a detailed chronology for Termination of the last cycle, but cannot date how the cycle began. Here we identify the central challenge in applying paleoglaciology to Quaternary glaciation cycles in general, and initiation of these cycles in particular. The challenge is the role of thickening sea ice in the Arctic Ocean, possibly becoming thick ice shelves floating over deep Arctic basins and high marine ice sheets grounded on shallow Arctic continental shelves. In that case, an Arctic Ice Sheet existed that was larger and less stable than the Antarctic Ice Sheet is now or in the past. Two approaches to this problem are presented as Part 1 and Part 2. In Part 1, the height of these former ice sheets is linked primarily to the strength of ice-bed coupling, which is deduced from glacial geology. It provides "snapshots" of ice sheets that give ice elevations independent of the past history and largely independent of ice-surface conditions, temperatures and the mass balance in particular. In Part 2, heights of former ice sheets depend on both their past history, particularly initial conditions, and changing surface conditions during the glaciation cycle. This provides a "motion picture" of ice sheets during the full cycle. The challenge is to establish a "mind meld" of these two approaches. They allow new perspectives on paleoglaciology.

1. Objective

Paleoglaciology explores the dynamics and history of former ice sheets. The Glacial Theory developed by nineteenth century naturalists saw dynamics and history as two sides of the same coin, using glaciology to determine the dynamics of existing ice sheets and glacial geology to determine the history former ice sheets (Andersen and Borns, 1994). Our objective is to restore the unity of glaciology with glacial geology. Unity was largely abandoned in the mid twentieth century when physicists began to quantify dynamic processes in ice sheets. This involved solving equations for the force, mass, and energy balance to obtain stresses resisting gravitational spreading of ice sheets, surface accumulation and ablation rates that control the size and shape of ice sheets over time, and temperatures within ice sheets that determine melting and freezing rates of basal ice and flow of basal water from sources to sinks. Inevitably, these solutions required tying the dynamics of ice sheets to atmospheric and oceanic dynamics and the dynamics of climate change. This emphasis linked ice sheets to the external environment at the expense of links to the bed under ice sheets, where glacial geology is produced and remains a record of the past behavior of ice sheets. This division between glaciology and glacial geology was highlighted in the 1970s. We review progress in glaciology and glacial geology with the goal of unifying these disciplines to obtain new perspectives on paleoglaciology.

Budd et al. (1971) developed a computer model to simulate steady-state dynamics of the Antarctic Ice Sheet. Their model calculated variations of stress and temperature with depth along ice flowlines from ice domes to ice margins using measured elevations, temperatures, and accumulation rates at the ice surface and known bed topography. Basal melting rates and temperatures below freezing were calculated for assumed values of the geothermal heat flux, and timelines along flowlines were calculated to help date oxygen isotope records and trace chemicals recovered from coreholes that serve as proxies for past climates. For a given geothermal heat flux, their model showed that seemingly minor variations in ice surface temperatures and accumulation rates could convert a bed that was largely frozen into a bed largely thawed. In addition, basal freezing rates were not calculated so the calculated basal melting rates would eventually float the ice sheet wherever

* Corresponding author.
E-mail address: terry.hughes@maine.edu (T.J. Hughes).

0277-3791/$ – see front matter
http://dx.doi.org/10.1016/j.quascirev.2013.08.023

the bed was thawed. Sugden (1977) subsequently corrected this defect in his application of the model to glacial erosion beneath the former Laurentide Ice Sheet. No attempt was made to include the dynamics of stream flow, even thought ice streams discharge up to 90 percent of the ice.

Denton and Hughes (1981), charged with reconstructing former ice sheets at the Last Glacial Maximum (LGM) for CLIMAP (Climate: Long-range Investigation, Mapping, and Prediction, a project of the International Decade of Ocean Exploration, 1970–1980), used glacial geology to determine former ice-sheet elevations along flowlines from the strength of ice-bed coupling deduced from the thawed fraction of the bed for slow sheet flow and the floating fraction of ice for fast stream flow. Their model had only a weak dependence on surface temperatures and accumulation rates and was independent of the geothermal heat flux. They presented two reconstructions of ice sheets at the LGM, maximum and minimum versions based on different interpretations of glacial geology, especially along marine margins of former ice sheets where glacial geology was below sea level and therefore both unmapped and undated. The dynamics of stream flow were crudely included simply by progressively reducing ice-bed coupling, with ice streams located using glacial geology.

A breakthrough in paleoglaciology was heralded when Hays et al. (1976) identified as "pacemaker of the ice ages" cycles of about 20,000, 40,000, and 100,000 years in precession of Earth's rotation axis with the equinoxes, in tilting of Earth's rotation axis toward and away from the sun, and in eccentricity of Earth's orbit around the Sun. These insolation variations controlled the intensity and distribution of solar energy over Earth's surface. This drew attention away from glacial geology as a primary data source for paleoglaciology, and focused attention on forcing variables which uniquely could be projected both into the past and into the future. This conclusion was premature. We now know that peak insolation and total insolation during the year at multiple north and south latitudes interact with environmental variables here on Earth to produce complex controls on glaciations.

Over the next decades, ice-sheet models based on the force, mass, and energy balance using grid points in the map plane instead of flowlines were made three-dimensional and time-dependent, and applied to both past and present ice sheets (see Huybrechts, 1990, 1994, l996; Huybrechts and T'siobel, 1995; and Huybrechts et al., 1991; for early examples of how this was done). At the same time, glacial geologists were using radiocarbon (Denton and Hughes, 1981; Chapters 1 and 2) and cosmogenic (Gosse and Phillips, 2001) dating techniques, and marine geology, to map and date former marine ice sheets in Antarctica (e.g., Anderson, 1999) and the Eurasian Arctic (e.g., Mangerud et al., 1999) on submerged continental shelves.

The new emphasis on external boundary conditions overlooked the most important fact of glaciology: the height of ice above the bed is determined primarily by the strength of ice-bed coupling, and almost equally important, the fact that glacial geology properly interpreted and dated shows how this strength of coupling changes through time.

As a demonstration, consider interior domes of both the Antarctic and Greenland ice sheets, with ice frozen to the bed, especially a rugged bed. This maximizes ice-bed coupling. Ice elevations lower as a frozen bed progressively thaws with distance from these ice domes, thereby reducing ice-bed coupling and allowing basal sliding of ice (Hughes, 1998; Chapter 3; Wilch and Hughes, 2000; Hughes, 2012; Chapter 24). Increasing the thawed fraction determines the amount of lowering, with ice above a thawed bed being about one-fifth lower than ice above a frozen bed (Denton and Hughes, 1981; Chapter 5). When a bed is completely thawed, further melting of basal ice increases the depth of basal water, so a thawed fraction of the bed becomes a floating fraction of ice where ice-bed uncoupling is complete (Hughes, 1992). Basal water will accumulate in low places of the bed, including river valleys eroded before the ice sheet formed, so a dendritic drainage system for basal water develops where ice-bed coupling effectively vanishes. These water drainage channels converge on larger depressions where fast stream flow replaces slow sheet flow because ice-bed uncoupling is linearized. This is shown dramatically by Rignot et al. (2011) for the Antarctic Ice Sheet using Radarsat-1 imagery. Up to 90 percent of Antarctic and Greenland ice sheets are discharged by these ice streams, much as most precipitation over continents reaches the sea via large rivers and their tributaries. Reduction of ice-bed coupling by increasing the floating fraction of ice along fast ice streams gives them a concave surface profile, whereas the surface profile for slow sheet flow is convex. These concave profiles dramatically reduce surface ice elevations (Hughes, 2012; Chapter 14). Ice streams typically end as low ice lobes on land or as floating ice shelves in water.

Apply progressive ice-bed uncoupling under ice sheets from interior ice domes to ice margins. Consider ice frozen to the bed under an ice dome 3000 m high, and over 3000 m thick for a marine ice sheet grounded below sea level. The ice surface will lower to 2400 m or more as the frozen bed thaws with increasing distance from the ice dome. Then a floating fraction replaces the thawed fraction where slow sheet flow converges to become fast stream flow. For marine ice streams, ice is typically about 1000 m thick at the ungrounding line where the ice stream becomes an ice shelf floating in the surrounding ocean. The buoyancy requirement of floating ice puts the ice elevation about 100 m above sea level. Further gravitational thinning can reduce the ice thickness to 300 m at the calving front, with ice 30 m above sea level. Ice initially 3000 m high has become only 30 m high, an elevation reduction of 99 percent, all due to progressive ice-bed uncoupling. A marine ice dome 3000 m high would itself eventually be only 30 m high, and be discerped by calving.

Glacial geology provides evidence for the strength of ice-bed coupling beneath former ice sheets. When an ice sheet 3000 m high at its center thins to 30 m high at its floating margin, the margin can be discerped by a variety of calving mechanisms linked to fracture mechanics and water-filled crevasses that produce a calving bay (Rist et al., 1996, 1999, 2002; Hughes, 2002; MacAyeal et al., 2003; Kenneally and Hughes, 2006). Can these calving bays migrate up the feeder ice streams and carve out the heart of the ice sheet? If so, ice sheets have the ability to self-destruct and thereby trigger rapid changes in global climate (Hughes, 1996, 2011). This is impossible for other components of Earth's climate system.

The possibility that ice sheets can self-destruct has a direct bearing on paleoglaciology because glacial geology contains records documenting destruction of former ice sheets. We attempt to show how glaciology and glacial geology address this possibility in two parts. Part 1 presents a "bottom up" strategy that emphasizes ice-sheet elevations controlled by the strength of ice-bed coupling deduced from glacial geology. Part 2 presents a "top-down" strategy that emphasizes how external environmental boundary conditions control how ice sheets change in size and shape through time.

In Part 1, ice-sheet elevations linked to the strength of ice-bed coupling features "snapshots" at specific times during a glaciation cycle of the Quaternary Ice Age. In this "bottom-up" approach, ice-bed coupling weakens when ice becomes thick enough to insulate the bed so a frozen bed thaws. Coupling is almost lost where basal water can drown the bed enough for slow sheet flow to become fast stream flow, and coupling vanishes under floating shelf flow. Glacial geology reveals these transitions (e.g., Hughes, 1998; Chapters 9 and 10; Kleman et al., 1999, 2008, 2010; Kleman and