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DATING GLACIAL LANDFORMS

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Definition

Dating glacial landforms. Applying geochronological tools (e.g., relative- and absolute-dating methods, etc.) to glacial landforms (e.g., moraines) to yield the timing of past glaciation (Moraine and Glacial Geomorphology and Landforms Evolution).

Introduction

Ever since scientists first recognized that glaciers and ice sheets were once larger in the past, they have desired to know the precise timing of past glaciation. Today, there is a more urgent need to tightly constrain patterns of past glaciation through time and space as projections of future global change rely upon knowledge from the past. Crude approaches have given way to complex techniques with increasing precision and decreasing uncertainty. Certainly, however, we are only a short way down a long path that carries us closer to a complete understanding and ability to date glacial landforms.

The techniques employed to date glacial landforms have been cleverly devised. For example, determining the growth rate of lichens and then measuring lichen diameters on moraine boulders to elucidate their exposure age (hence the timing of moraine deposition), using the patterns of tree-ring thickness to "cross-date" exhumed stumps that were eroded by a former glacier advance, and measuring the accumulation of isotopes in rocky surfaces that result from the bombardment of cosmic radiation to calculate the time elapsed since glaciers retreated. Our current understanding of when former glaciations occurred is better than ever, but is far from complete. Even with the dating techniques currently available, we could vastly improve our current understanding with more resources and time.

Here, I focus on dating glacial landforms, such as moraines and outwash terraces (depositional landforms), and glacially-eroded bedrock features and U-shaped troughs (erosional landforms). Thus, not included are the wide variety of techniques used to date stratigraphic sequences of glacial sediments. In some cases, the boundaries of dating glacial landforms and dating glacial sediments are blurred. For example, moraines comprise glacial sediments, and dating sediments associated with a landform can constrain landform age. However, focus on dating landforms inherently results in omitting certain landforms from this entry whose age in absolute time can only be constrained by dating glacial stratigraphy, such as subglacial depositional landforms (e.g., eskers) or depositional/erosional landforms (e.g., drumlins). Of course, dating both landforms and glacial sediments, combined with additional information, such as records of global ice volume from $\delta^{18}O$ measurements from benthic marine organisms, has led to the present understanding of the timing of Earth's glaciations.

The focus on dating glacial landforms inherently results in discussion of recent (middle and late Quaternary) landforms. In some cases, landforms survive from pre-late Pleistocene glaciations, which can be a result of slow rates of landform degradation and the survival from erosion from successive glaciations. However, in most cases where old landforms remain intact on the landscape (e.g., pre-late Pleistocene), the ability to date them is hampered by the limits of the method or by the increasing uncertainty of dating techniques back in time. Finally, this entry focuses on the dating tools that are widely used today. Although some commonly used relative- and

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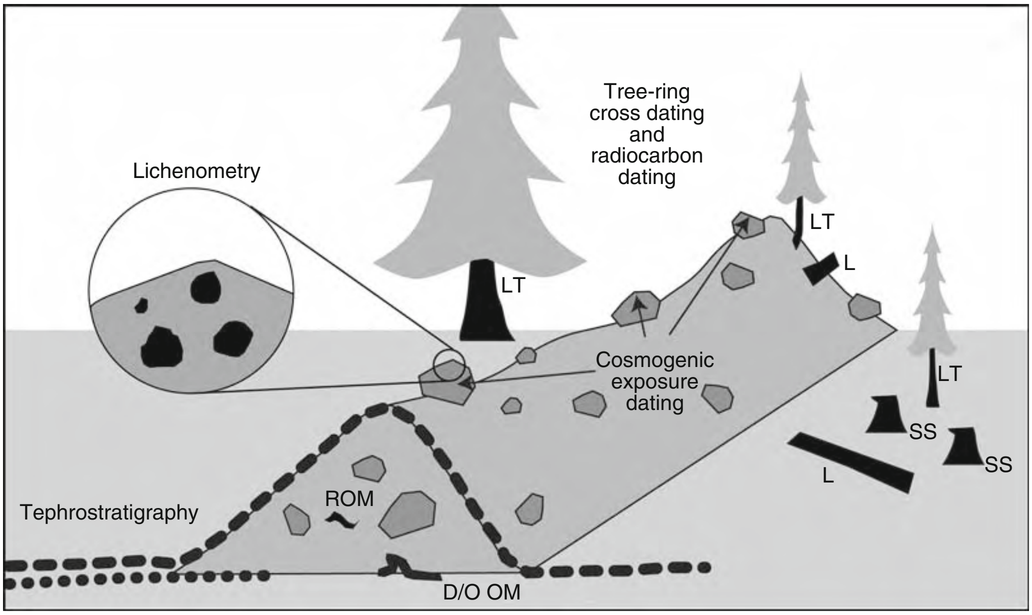


Dating Glacial Landforms, Figure 1 Common methods used to date moraines. Targets for radiocarbon and tree-ring cross dating are labeled: *LT* living tree, *L* log, *SS* sheared stump, *ROM* reworked organic material, *D/O OM* deformed/overridden organic material. *Dashed* and *dotted lines* represent tephra layers.

calibrated-dating techniques are discussed, more emphasis is placed on the absolute-dating techniques that are most commonly applied to glacial landforms (Figure 1).

Relative-dating techniques

The initial and most fundamental approach to dating glacial landforms is ordering landscape features in a relative sense. In terms of moraines, those closer to the ice source are younger because in almost all cases subsequent glaciations obliterate prior surface features beneath their footprint. When dealing with moraines deposited by alpine glaciers (i.e., moraines in mountain valleys), assigning relative ages to moraines is fairly straightforward. Much like assigning relative ages to rock layers (stratigraphy), there is a relative age assignment to surface features (morphostratigraphy). In some cases, moraines can be cross-cutting, where younger moraines are deposited on top of, and truncate, older moraines. A classic example is Bloody Canyon, eastern Sierra Nevada, USA (see Phillips et al., 1990), but there are a surprising number of other examples.

In rare cases, for example where polar ice sheets are cold-based, the obliterative nature of glacier flow is replaced by non-erosive characteristics. In these settings, it is possible to preserve glacial landforms that were formed during previous glacial cycles, and the relative ordering of landforms is more complicated. On the other hand, this rarely happens in alpine landscapes. In other cases, subglacial bedforms (e.g., drumlins, megaflutes, etc.) in areas that were occupied by Pleistocene ice sheets reveal shifting flow directions. The preservation of these stacked sequences of bedform orientations reveals that, at least in some locations, bedforms can be preserved from not just the most recent flow direction.

Slightly more sophisticated approaches to the relative dating of glacial landforms rely on the physical and chemical weathering that takes place on and within glacial deposits. The application of soil chronosequences to moraine and outwash surfaces has been used to assess the relative age of these features, and in some cases to correlate glacial landforms from valley to valley in a given mountain range. In particular, soil thickness, B-horizon thickness, B-horizon development, and weathering-rind thickness measured in clasts in soil profiles have been used as indicators of relative age (Porter, 1975; Burke and Birkeland, 1979; Colman and Pierce, 1986; Birkeland et al., 1991). The weathering of surface rocks has also been employed as a relative-age indicator, specifically, characteristics such as the abundance and depth of pitting, grussification, and hardness have been used to make relative subdivisions of moraines (e.g., Birkeland et al., 1979). The degree of degradation of depositional landforms (landform morphology) has also been used as a relative-dating parameter. Because moraines are originally deposited with relatively steep slopes that degrade with time, the steepness of moraine slopes, or degree of surface roughness within hummocky moraine belts that we see on the landscape today is partly of function of moraine age. Slope angle, crest width, and the degree of gullying are parameters that have been measured and linked with relative age (e.g., Kaufman and Calkin, 1988).

Despite the factors that complicate the accuracy of relative-dating techniques, they nonetheless remain useful. Relative-dating techniques are useful for correlating moraines from valley to valley, provide the only chronology in many cases where materials for absolute dating are absent, and act as an aid even when absolute dating is available. Furthermore, it is generally less time-consuming and less expensive to employ relative-dating techniques versus absolute-dating methods, and thus characteristics of many more moraines can be included in a dataset. Finally, because of the high cost of many absolute-dating techniques, using relative-dating methods to correlate a low number of landforms with absolute-age control to many more landforms of the same properties across a region is a powerful approach.

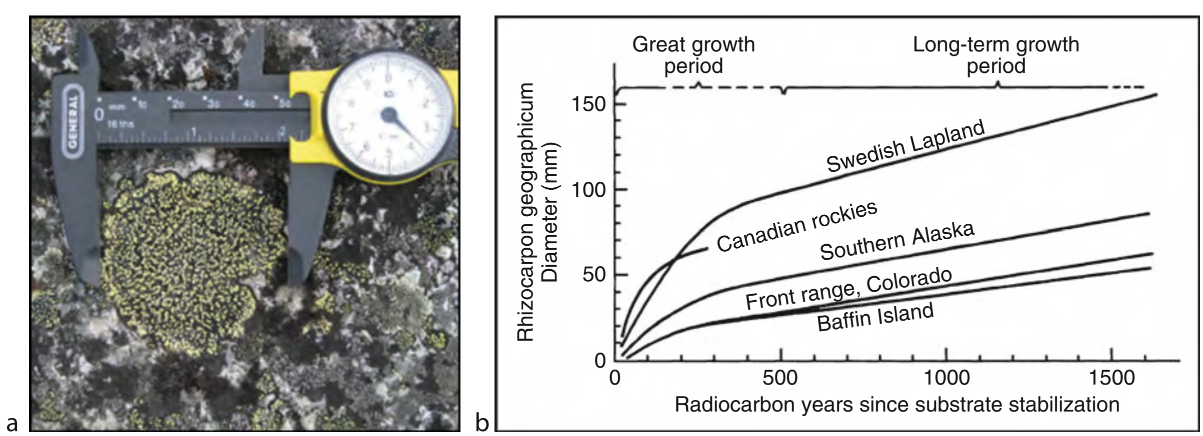
Lichenometry

Lichenometry is a surface-exposure dating method that uses lichen-growth rates to infer the age of young (few thousand years old or younger) glacial landforms, typically bouldery deposits such as moraines. The technique combines measurements of the size of lichens growing on rocky glacial deposits with independently derived lichen growth rates to derive lichen age, and thus moraine age. Lichen types that grow radially and regularly are used, most commonly the crustose lichen genus *Rhizocarpon*, where *R. geographicum* is specifically targeted in most cases, but field identification to the species level is difficult (Figure 2). The method has been widely applied since its development in the mid-twentieth century (Beschel, 1950).

Several approaches have been employed to measure lichens, and a distinct advantage of lichenometry over other techniques is its simplicity (Bradwell, 2009). One approach is to measure the diameter of dozens to hundreds of semi-circular lichens on boulders scattered about on a moraine surface. The largest diameter measured, or the average of the five largest diameters measured, can be used with the growth curve to obtain a surface age. Additional approaches include determining the size frequency of all lichens in a representative area, or measuring the total lichen cover on a substrate. Ongoing research includes more advanced statistical approaches to extract the most meaningful age from field measurements (cf. Bradwell, 2009) and placing more emphasis on the controls on the pattern of lichen growth (Loso and Doak, 2006).

Used in the right circumstances, lichenometry can be a successful numerical-dating method. This success relies on several factors. First, because lichen-growth rates are a function of a variety of conditions (e.g., regional climate, substrate, microenvironment, species competition), the growth rate needs to be well constrained in the area of application (Figure 2). Independently dated surfaces upon which lichens grow that are commonly used for calibration include tombstones, mining waste, and archaeological sites. Furthermore, because growth rate is not linear, but rather has an initial period of fast growth, followed by a slower, linear growth rate, multiple calibration points are required to quantify the growth rate through time (Figure 2). With a well-constrained growth rate that spans the same time period of study and that was measured on similar substrates, the lichenometry method can be reasonably accurate. Because of increasing uncertainty back in time, changing climatic conditions through time, and the intersection of individual lichens as surfaces become heavily colonized, the lichenometry method becomes less reliable beyond a few thousand years. The highest accuracy with lichenometry is its application to glacial landforms deposited within the last millennium.

There are many uncertainties with lichenometry. Many involve how applicable the growth curve is to any single area of study. As with any surface-exposure dating technique, lichenometry best dates the timing of landform stabilization, as in the case of moraines (Putkonen et al., 2008). Furthermore, there is an unknown amount of time it takes for lichens to colonize a surface, but this is thought



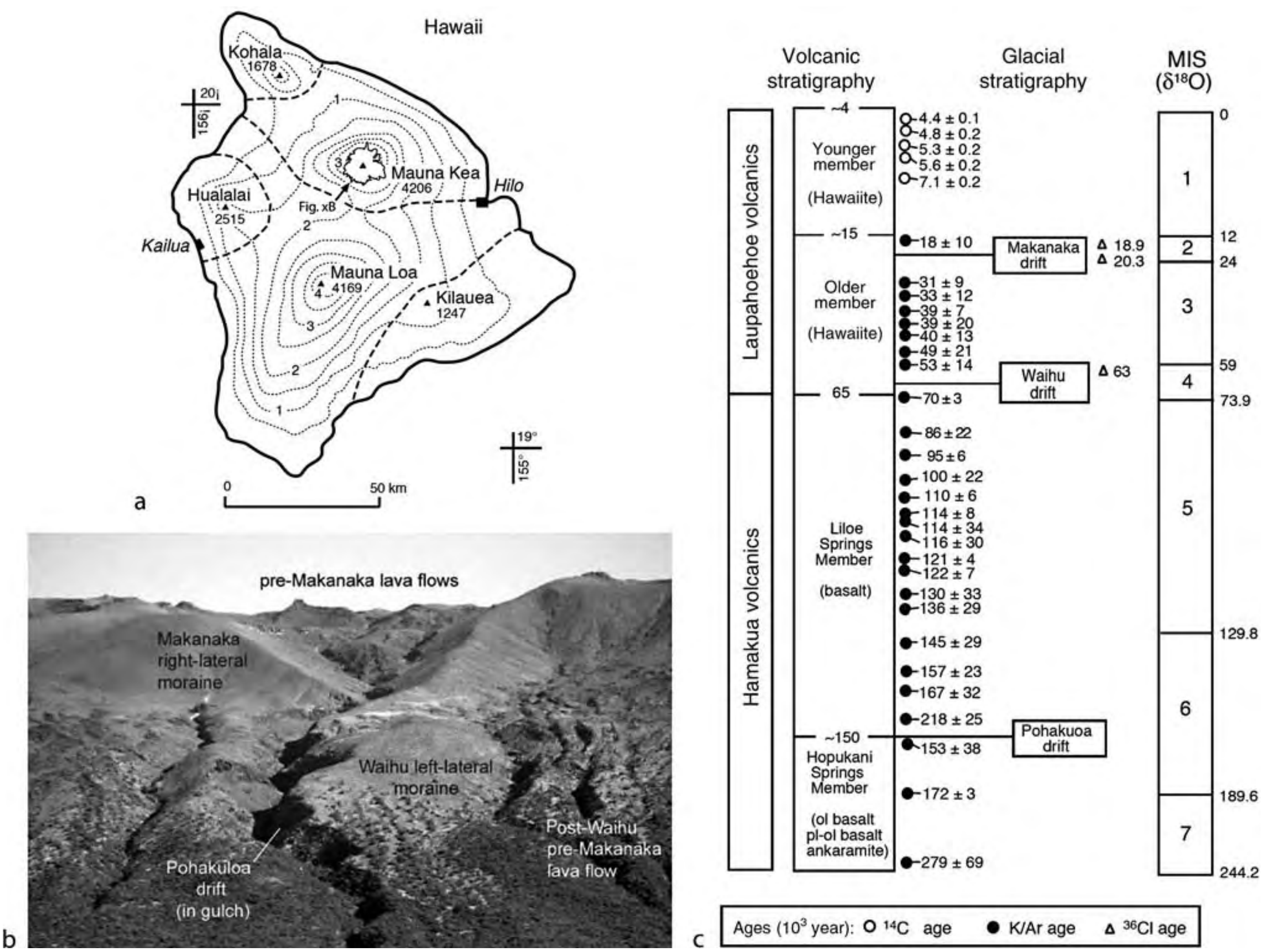
Dating Glacial Landforms, Figure 2 (**a**) Photograph of *R. geographicum* thallus measured on a moraine boulder (from Young et al., 2009). (**b**) Selected *R. geographicum* growth curves (from Calkin and Ellis, 1980).

to be years to decades, not centuries. Finally, it should be mentioned that even where well-constrained lichen growth curves are absent, lichenometry can still serve as a valuable relative-dating technique. For example, determining the largest lichen diameters on individual moraines among a sequence can be useful where correlating moraines between valleys across a mountain range. Additionally, the full range of lichen diameters present on a large group of moraines might indicate whether all the moraines are of similar age, or whether moraine ages span millennia (Solomina and Calkin, 2003; Barclay et al., 2009a).

Association with volcanic deposits

In regions where glaciers exist in proximity to volcanoes, glacial landforms can be dated by their association with volcanic flows that are radiometrically dated. In addition, where glaciation takes place near volcanic arcs or other areas with explosive volcanism, the age of glacial landforms can be constrained by tephrostratigraphy. These techniques are useful throughout the Quaternary, spanning the time period represented by the record of glacial landforms.

One example of dating glacial landforms by their association with dated volcanic flows took place near the summit of Mauna Kea, Hawaii (Figure 3). Mauna Kea periodically supported an ice cap during the Pleistocene, from which outlet glaciers flowed radially part way down the volcano flanks. Potassium–argon dating of lava flows that overlie and underlie drift units associated with two of the most recent moraines deposited on Mauna Kea provided bracketing ages on moraine age (Porter, 2005). A second example regards dating moraines in Patagonia



Dating Glacial Landforms, Figure 3 Example of dating volcanic deposits to constrain moraine age from Hawaii modified from Porter (2005). (**a**) Hawaii showing location of Mauna Kea. (**b**) Oblique photograph showing late Pleistocene moraines deposited on flank of Mauna Kea. (**c**) Summary diagram showing K/Ar ages on lava flows interbedded with glacial drift units; in particular the K/Ar ages constrain the deposition of the Waihu and Makanaka moraines. Note how K/Ar ages compare to ^{36}Cl exposure ages from boulders on the moraine surface.

bounded by basalt lava flows. Argon–argon and potassium–argon dating was used to constrain the age of several moraines deposited during the middle and late Pleistocene (Singer et al., 2004). In some locations (e.g., Iceland), volcano-glacier landforms occur, such as table mountains (or tuyas), which are volcanoes that erupt subglacially and eventually emerge through the ice, ending in a subaerial eruption phase. These features can be dated radiometrically using volcanic materials (Kaufman et al., 2001), or by cosmogenic-exposure dating (see below) on the subaerial surface lava flows (Licciardi et al., 2007).

Volcanic ash, or tephra deposits, that lie beneath or overlie moraines can be used to constrain moraine age. Tephrostratigraphy is the field of matching unknown tephras at a study site with a database based on prior work of tephras of known ages and chemistries. By analyzing tephra chemistry at a study site and comparing it to a database, tephras and their ages can be identified. Because tephras are airfall deposits, they form stratigraphic layers on the Earth's surface. Thus, they can underlie or overlie glacial landforms like moraines, and provide maximum and minimum ages, respectively (Figure 1). In some cases, tephra can be reworked into glacial depositional landforms, which can also provide a maximum age of moraine deposition. In this way, moraines can be dated by their association with tephras. Some examples are dating moraines in Alaska (e.g., Begét, 1994), and in the Cascade Range, western U.S. (e.g., Porter, 1976; Heine, 1998).

Radiocarbon dating

Radiocarbon dating is probably the single most important dating method in Quaternary science. With a usable age range between ~300 and ~40,000 years in most applications, and sample types that include most organic-material, radiocarbon dating is widely used to date sediments and landforms. Typical analytical uncertainties are ~2–5%, although uncertainty typically becomes larger when ages are calibrated into calendar years. In terms of dating glacial landforms, radiocarbon dating has been most useful for dating moraines, although ultimately the dating of the moraine itself arises by dating the sediments that comprise the moraine, or sediments associated with moraines (Figure 1). Studies have applied radiocarbon dating to sediments below, within, and above moraines to provide maximum (below and within) and minimum (above) age constraints.

Dating moraines with radiocarbon works best in environments where glaciers flow across forested landscapes. Radiocarbon dating of wood is used to constrain moraine age in several ways. When a glacier advances into a forest, its proglacial sediments sometimes partly bury tree trunks. By the time the glacier snout reaches these partly buried trees, it shears them off part way up the tree trunk. Following moraine deposition and glacier retreat, the in-situ sheared tree trunks become exhumed and available for sampling. Radiocarbon dating of their outer rings provides a maximum age on moraine formation. In other cases, sheared trees become incorporated in moraine sediments, become deposited on the landscape as erratic logs, or are deposited in till blanketing the landscape, and eventually exposed or washed out. In these cases, radiocarbon dating outer rings of glacially transported logs also provides maximum ages on moraine formation. There are many good examples of using radiocarbon dating to date Holocene moraine formation from southern Alaska (e.g., Barclay et al., 2009a), western Canada (Menounos et al., 2009), and in the Alps (Joerin et al., 2006). In rare cases, glacier snouts tilt trees as they deposit moraines; in this case, radiocarbon dating the damaged tree can provide a close-limiting age on moraine formation. Where logs are involved, radiocarbon dating is commonly used in conjunction with tree-ring cross dating, which is a more precise method to constrain moraine age (see below) (Figure 4).

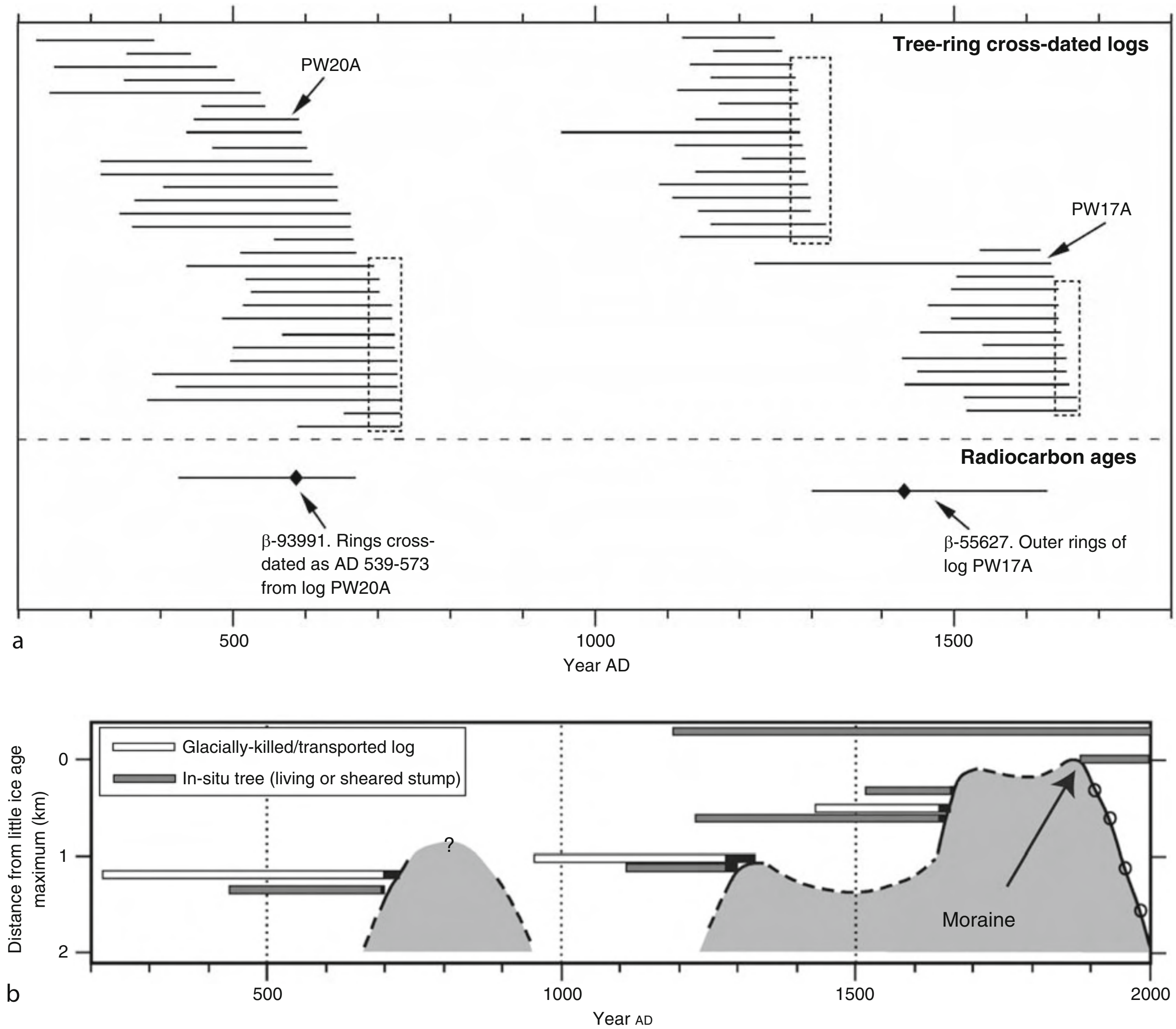
There are many cases, of course, where entire logs are not preserved, but rather pieces of wood or cones and other organic material are reworked and incorporated into moraine sediments. In places where glaciers advance through marine embayments and fjords, they can rework fossiliferous marine sediments into their moraines. In Greenland, for example, radiocarbon dating of reworked fossil material, including marine bivalves, whale bones, and even a walrus tusk provided maximum constraints on the age of late Holocene moraine formation (Weidick et al., 2004). In some cases, glaciers overrun lakes or bogs and deform peat; in these cases radiocarbon dating of the uppermost layers of disturbed peat provide close maximum age constraints on moraine formation (Mercer and Palacios, 1977; Buffen et al., 2009).

In most cases, finding organic materials below or within moraines is rare. More widely used are organic deposits that accumulate behind or on top of moraines. Obtaining so-called basal radiocarbon ages from lakes and bogs is a powerful approach to providing minimum constraints on moraine age in both alpine and continental ice sheet settings (e.g., Thackray et al., 2004; Lowell et al., 2005).

Radiocarbon dating of glacial landforms includes features in addition to moraines. Organic deposits on top of, beneath, and incorporated into outwash deposits allow radiocarbon dating to constrain the age of outwash terraces (e.g., Porter et al., 1983; Hamilton, 1986). Radiocarbon dating has also been used extensively to date raised glaciomarine landforms, such as ice-contact deltas, that are deposited during ice retreat in isostatically recovering (emerging) landscapes (Dyke, 1999). In these cases, radiocarbon ages of in-situ bivalves from the delta sediments provide a direct age on ice-contact delta formation.

Tree-ring cross dating

Tree-ring cross dating is a precise means to date logs, and where logs are associated with moraines (see above), tree-ring cross dating can provide more precise age control than radiocarbon dating alone (Wiles et al., 1996).



Dating Glacial Landforms, Figure 4 Example of using radiocarbon and tree-ring cross dating. (**a**) *Horizontal black lines* in upper portion of figure represent the lifespan of each cross-dated log found in front of a retreating glacier in south Alaska; *dashed line boxes* represent timing of glacier advance when the majority of the trees were killed, indicating glacier advance. The lower portion shows the 2-σ age range (*black bar*) and median (*diamond*) of radiocarbon ages of the outer rings of select logs. Modified from Barclay et al. (2009a). (**b**). Time-distance diagram of the Tebenkof Glacier, south Alaska, constrained by tree-ring cross dating of glacially transported trees and in-situ sheared stumps that were killed upon glacier advance and living trees growing on and outboard of the Little Ice Age moraine. Modified from Barclay et al. (2009b).

The principles rely on wiggle-matching patterns of tree-ring widths between a "master" tree-ring width series and tree rings in a specimen of which the age is unknown. In this manner, logs can be placed in absolute time, and their age of death in some cases can be constrained to a single year. Thus, over time periods spanned by a master ring-width series, using tree-ring cross dating to determine the age of trees can be much more precise than radiocarbon dating.

Because of regional variations in climate and other factors that influence tree growth, tree-ring master series are constructed and applied within specific regions. Because living tree-ring width time series only go back a few centuries in time, subfossil trees are used to extend the time series farther back in time. Once master chronologies exist, placing glacially transported logs and sheared stumps into an absolute age provides tight age control on moraine formation. For example, when glacially-transported logs in glacier forefields are dated by cross dating, they reveal when a stand of trees were glacially overrun, and hence provide a maximum age for moraine formation (Barclay et al., 2009b). Similarly, in-situ stumps that have been sheared from advancing glaciers also provide maximum ages of moraine formation, and also pinpoint the exact location of glacial overriding (Barclay et al., 2009b; Menounos et al., 2009) (Figure 4). Tree-ring counting from living trees that grow on moraines can provide a close minimum age of moraine formation, but the method can be complicated somewhat by the ecesis time (the time it takes for a tree to germinate on fresh glacial deposits). In some cases, glaciers that advance into forests can damage or tilt trees in the processes of moraine deposition. In these cases, cross-dated damaged trees can provide a precise time of moraine formation (Wiles et al., 1996; Barclay et al., 2003).

Cosmogenic-exposure dating

Cosmogenic-exposure dating (also referred to as surface-exposure dating and terrestrial cosmogenic-nuclide dating) has emerged over the last two decades as the premier chronological tool to date glacial landforms such as moraines, erratic boulders, and glacially eroded bedrock. As with the other dating methods above, the details of the method are not discussed herein, but rather focus is placed on its application to dating glacial landforms. See Gosse and Phillips (2001) for a thorough treatment of background and fundamental principles.

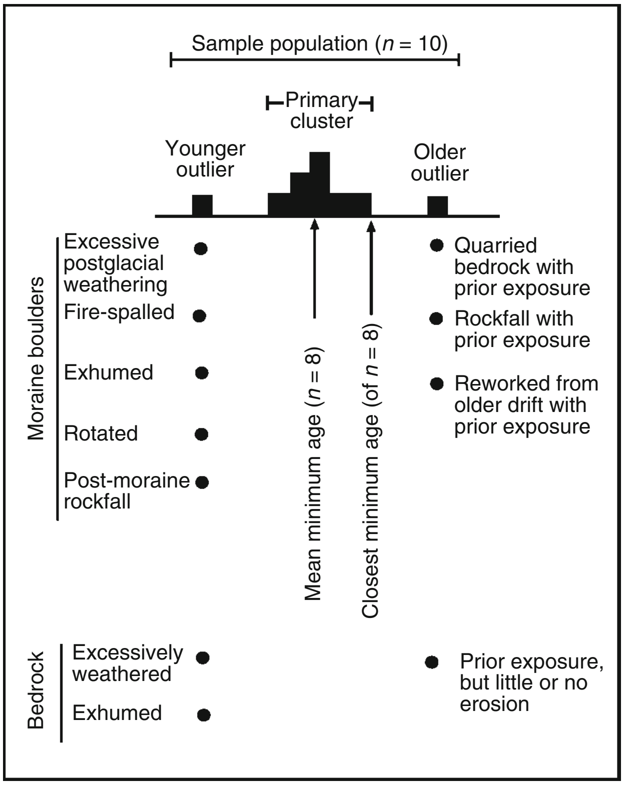
Briefly, a family of isotopes is produced in Earth's rocky surface (mostly in the upper few meters) as a result of cosmic-ray bombardment. Many of the resulting "cosmogenic" isotopes are radioactive (e.g., ^{14}C, ^{10}Be, ^{26}Al, ^{36}Cl), whereas others are stable (e.g., ^{3}He, ^{21}Ne). Some of the isotopes are produced solely from cosmic ray bombardment (e.g., ^{10}Be), whereas others can be formed by additional means (e.g., ^{36}Cl). The different isotopes have differing target, or parent, minerals. For example, applications of ^{10}Be mainly use quartz-bearing rocks, whereas applications of ^{3}He commonly rely on olivine phenocrysts in igneous rocks. And, some methods use a "whole-rock" approach because there are several parent elements that produce the cosmogenic isotope (e.g., ^{36}Cl). Furthermore, the radioactive cosmogenic isotopes (radionuclides) are used for different applications depending on their varying half lives.

Because organic matter and glacial deposits are rarely associated, radiocarbon dating cannot be applied in many areas. On the other hand, cosmogenic-exposure dating can be used in more widespread locations because the target materials are rocky deposits typical of glacial landscapes. Moraines, more specifically boulders on moraine surfaces (Figure 1), have been the primary target of cosmogenic-exposure dating, although there have been many other applications. Ultimately, because the target samples (e.g., boulders and bedrock) for cosmogenic-exposure dating are different, and perhaps more common, from organic material required for radiocarbon dating, and are more common than materials targeted for other dating techniques mentioned above, cosmogenic-exposure dating has revolutionized the ability to date glacial landforms. Furthermore, although radiocarbon dating is usable only to about 40,000 years ago, the analytical limit of cosmogenic-exposure dating extends back hundreds of thousands to millions of years.

Early studies that applied cosmogenic-exposure dating to moraines and glacial boulders took place primarily in the western U.S. and Antarctica (e.g., Phillips et al., 1990; Brook et al., 1993). Since then, there have been dozens of studies that have generated moraine chronologies from around the globe. These studies range from moraines deposited by mountain glaciers in tropical latitudes (e.g., Smith et al., 2005a), middle latitudes (e.g., Phillips et al., 1997), and high latitudes (e.g., Briner et al., 2005a). Moraines deposited by ice sheets have also been targeted (e.g., Balco et al., 2002). Until recently, most research has focused on late Pleistocene deposits (~120,000–11,700 years ago), and most specifically, on deposits created during the peak of the last glaciation (~25,000 to ~11,700 years ago). This time period has the advantage of being geologically young enough such that moraines are not too intensely weathered and degraded, yet old enough that concentrations of cosmogenic nuclides are high enough to be analytically measurable. In some locations with apparently slow rates of erosion and moraine degradation, exposure dating works reasonably well farther back in time (Smith et al., 2005b; Licciardi and Pierce, 2008; Porter and Swanson, 2008), but in general, moraine chronologies become increasingly scattered beyond 50,000–100,000 years ago. Recently, as analytical techniques and chemical procedures in preparation laboratories have improved, dating moraines deposited during the Holocene (last 11,700 years), and even during the last millennium, has become more common (Schaefer et al., 2009; Licciardi et al., 2009).

Producing reliable moraine chronologies with low uncertainty relies on several important criteria (Figure 5). To ensure that boulders receive constant exposure to cosmic-ray bombardment since their deposition, the upper, skyward-facing surfaces of large boulders (usually >1 m high, but even larger boulders typically are more stable on moraines) are sampled. Because there are a number of geologic and analytical uncertainties that can affect any given boulder age, researchers typically average cosmogenic-exposure ages from five to ten boulders per moraine crest. Although analytical uncertainties can be significant (particularly for young surfaces), geologic uncertainties, and uncertainties in isotope production rates and spatial scaling rules are typically more significant. Because this entry emphasizes the application, the discussion herein is limited to the most important geologic uncertainties.

Because moraines are depositional features, gravity efficiently decreases moraine slope angles and thus moraines degrade through time (Bursik, 1991). The tilting and rolling of surface boulders, and the exhumation of subsurface boulders to the surface, yield cosmogenic exposure ages that are younger than the true timing of moraine deposition (Figure 6). Thus, moraine degradation severely hampers tightly-clustered moraine chronologies, and skews average ages toward a younger age. In addition

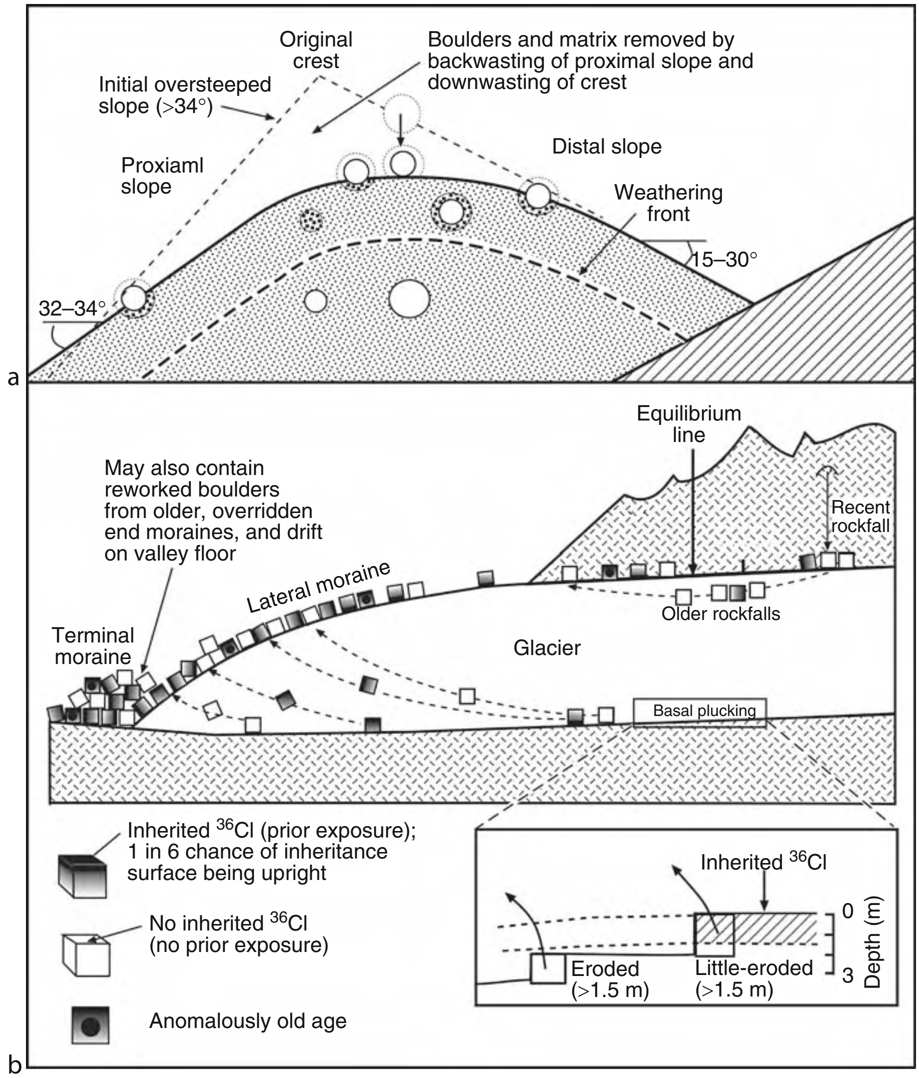


Dating Glacial Landforms, Figure 5 Complicating factors that potentially influence the distribution of cosmogenic exposure ages of moraine boulders (*top*) and glacially-eroded bedrock (*bottom*). In addition, the diagram provides guidelines for interpreting a population of cosmogenic exposure ages of moraine boulders, such as identifying outliers and interpreting an age cluster as the mean minimum age versus closest minimum age. From Porter and Swanson (2008).

to moraine degradation, rock-surface erosion acts upon boulder surfaces and is a process that decreases the cosmogenic-isotope inventory, and thus decreases cosmogenic-exposure age. Despite rock-surface erosion being most important for poorly indurated lithologies and/or for pre-latest Pleistocene moraines, it is commonly difficult to determine how much rock-surface erosion has taken place, and thus remains a potential complication (Figure 6). Sampling boulder surfaces that retain glacial striations or polish avoids this issue, but these features are rarely preserved. Multiple cosmogenic isotopes with different depth-production profiles have also been used to solve for surface erosion (Phillips et al., 1997). Finally, partial burial of moraine boulders by loess or seasonal snowfall also yields cosmogenic exposure ages that are younger than the timing of moraine deposition. Correcting for snow and loess cover is challenging because it is difficult to reconstruct the magnitude of shielding for the duration of a sample's surface history. Most researchers collect samples from tall boulders on high parts of the landscape to increase the likelihood that boulder surfaces are windswept. In some cases, collecting samples during the snow season to evaluate modern snow cover is helpful. In any case, these three complications (moraine degradation, boulder-surface erosion, snow/loess shielding) lead to ages younger than the true timing of moraine formation, which further complicates determining their relative importance. In all cases, careful field sampling can significantly reduce these complications.

Another type of geologic uncertainty arises from isotopic inheritance (Figure 6). Isotopic inheritance refers to cases where moraine boulders are deposited on a moraine surface already containing some inventory of cosmogenic isotopes. These isotopes are "inherited" from a period of prior exposure, a situation that results from inefficient glacial erosion, rockfall onto a glacial surface, or reworking of older materials. In cases where eroded and transported blocks are deposited with their previously exposed surface facing skyward, a chance exists that inherited isotopes combine with isotopes that accumulate during the targeted period of exposure. Thus, these boulders yield cosmogenic exposure ages that pre-date the actual time of moraine deposition. In many cases, inheritance is a much less likely complication than moraine degradation. However, certain contexts exist in which the likelihood of inheritance occurring significantly increases. Primarily, inheritance occurs where glaciers are cold-based or polythermal, and thus do not deeply erode landscapes. There are several examples at high latitude study areas of the prevalence of inheritance and the complication that it imposes in producing moraine chronologies (e.g., Briner et al., 2005b). Indeed, in many settings the presence/absence of isotope inheritance has been used to elucidate patterns of glacial erosion (Sugden et al., 2005; Briner et al., 2006). Another likely context in which inheritance can be common is in short glaciers (e.g., cirque glaciers) where the proportion of head- and valley-wall-sourced boulders is relatively high compared to subglacially sourced boulders. Finally, in some cases, very near the terminus of glaciers and ice sheets, erosion becomes less efficient, and reworked boulders have a better chance of appearing in younger deposits (e.g., Balco et al., 2002).

Assessing the various geologic means by which scatter in a population of cosmogenic-exposure ages arises is critical to ultimately assigning moraine age (Figure 5). To date, there have been several investigations regarding the relative importance of these major complicating factors (Hallet and Putkonen, 1994; Putkonen and Swanson, 2003). In most cases, these studies document the stronger and more typical influence of moraine degradation on a suite of cosmogenic-exposure ages. For this reason, many researchers interpret cosmogenic-exposure ages as providing the timing of moraine abandonment and hence the onset of deglaciation following a glacier advance. That is, boulders that are sampled at the moraine surface were the last deposited on the moraine crest just prior to retreat of the ice margin. In addition, due to the prevalence of moraine degradation, many researchers interpret cosmogenic exposure ages as minimum ages for the onset



Dating Glacial Landforms, Figure 6 (**a**) Diagram illustrating complication of applying cosmogenic exposure dating to degrading moraines, in particular, the issue of boulders exhumed to the moraine surface as the crest lowers through time. The additional complication of boulder erosion is also illustrated; "zoned" boulders indicate weathering rinds and grussification. (**b**) Sketch showing different processes that lead to inherited cosmogenic isotopes (e.g., ^{36}Cl) in moraine boulders. Modified from Porter and Swanson (2008).

of deglaciation. In some rare cases where moraine crests are stacked together, the outermost moraine crest may not date the onset of deglaciation (Gosse et al., 1995). A misconception that still arises in the literature is that cosmogenic-exposure ages yield the timing of glacier advances, or the timing of glacier maxima. This can result in incorrectly assigning leads/lags between one study area compared with glacier and climate-change records from elsewhere. Thus, it is critical to have a firm understanding of complications in producing cosmogenic-exposure dating chronologies, and to acknowledge what part of the glacial cycle the ages constrain.

In addition to moraines, cosmogenic-exposure dating has been applied to other glacial landforms with great success. Common glacial landforms targeted with cosmogenic-exposure dating are outwash and marine terraces (e.g., Anderson et al., 1996). Because fine-grained sediments (as opposed to large boulders on moraine surfaces) can become mixed after their deposition by soil and bioturbation processes, merely dating a collection of surface pebbles potentially would yield an age younger than the true timing of sediment deposition. An additional complication is that a small proportion of inherited isotopes can be present in sediments. Both of these complications can be elucidated by collecting samples of sediment along a depth profile, measuring cosmogenic isotope concentrations in each, and matching the resulting profile of concentration versus depth to the expected theoretical profile. In this manner, inheritance can be detected (if, with depth, concentrations decrease to a value above zero, which represents the inherited component). Furthermore, the depth profile can be extrapolated to the surface to estimate the surface exposure age. Although useful, this approach does require multiple measurements for a single surface age.

Dating glacial erosional landforms

Glacial-erosional landforms, such as the variety of bedrock landforms common in glaciated landscapes (e.g., roche moutonnées, whalebacks, etc.), are commonly polygenetic features (depending on their size). In many cases, these bedrock features have been reshaped during the most recent period of glaciation. Cosmogenic-exposure dating has been employed with success to assign ages to glacially eroded bedrock surfaces. Dating the landform, in most cases, is accomplished to place timing on the most recent period of ice recession. Many studies have successfully placed constraints on ice sheet thinning (e.g., Stone et al., 2003) and recession (e.g., Briner et al., 2009) using transects of samples from glacially eroded bedrock landforms.

In some ways, cosmogenic-exposure dating can also constrain the actual age of erosional landforms. For example, where cosmogenic-isotope inheritance exists, the magnitude of glacial reshaping of the landform during the last glacial cycle can be determined to be minimal. In fjord landscapes, the spatial distribution of isotope inheritance can constrain how much these large-scale glacial landforms evolved during the most recent glaciation (e.g., Sugden et al., 2005; Briner et al., 2006). In some cases, researchers have used two isotopes with different half lives (e.g., ^{10}Be and ^{26}Al) to assess glacial erosion not only for the most recent period of glaciation, but the general magnitude of erosion and landscape modification during prior glaciations (e.g., Bierman et al., 1999). An additional tool applied to determine the timing of formation of large-scale glacial erosional features, such as fjords, is thermochronology. In particular, low temperature thermochronology techniques (e.g., (U-Th)/He and ^{4}He/^{3}He; Reiners and Brandon, 2006) have recently placed constraints on the evolution of glacial landscapes and the timing of major fjord incision (Shuster et al., 2005).

Summary

Reliably, dating glacial landforms are fundamental to the field of paleoclimatology. All dating methods have uncertainties and room for improvement. Just as multiple proxies are needed to reconstruct past climate change, or multiple climate models should be used to predict future climate change, more than one geochronological tool should be employed wherever possible to date glacial landforms. Newer and more powerful techniques (e.g., cosmogenic-exposure dating) need to be used in concert with relative-dating techniques, beginning with a foundation of geomorphic mapping and morphostratigraphic relationships. The geochronologic toolbox employed to date glacial landforms is expanding, and it is exciting to think about what additional techniques will be available in the future. At the same time, however, the tools already in hand are sufficient to read landscapes upon which there are widespread traces of former glacier change.

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DEAD ICE

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Definition

The dead ice is defined as the ice which does not move, that is, it becomes stagnant. Generally, dead ice is formed due to detachment of/from the active glacier and topographic conditions do not allow for its movement, and it is covered with thick piles of moraine/debris, which act as insulator and protect from quick melting. It is also known as stagnant ice.

DEBRIS

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Synonyms

Boulder; Detritus; Scree material

Definition

Unconsolidated sediment, larger than 1 mm, of angular or rounded angular fragments of boulders (clasts), predominantly originating from physical weathering.

Introduction

Debris occurs in a wide range of environments, but most significantly in those where physical weathering rates dominate, such as periglacial environments. Hence debris is common in glacierized areas, affecting glaciers and ice in the ground, as well as their appearance and characteristics, in a great variety of ways. Debris e.g. significantly influences the glacier melt and the glacier movement. Debris can be found on the surface of glacier ice (supraglacial), within glacier ice (englacial), as well as below (subglacial) and beside glacier ice. Coarse debris in periglacial scree slopes plays an important role in the development of so-called rock glaciers (Barsch, 1996, Haeberli et al., 2006). The focus here is on debris related to glacial environments.

Debris sources

The principal sources of debris are mass movements such as rockfalls, rock avalanches, debris flows, debris-laden ice, and snow avalanches from surrounding mountain slopes (Hambrey et al., 2008; Kirkbride, 1993). These sudden and sometimes vast events of debris relocation are common phenomena in alpine regions. The amount of debris that reaches a given glacier depends on the characteristics and extent of the catchment area, especially its weathering and erosion rates (Haeberli, 1986). These again are influenced by the lithology of the source material. Scree slopes (debris deposits) situated in a periglacial environment favor the development of firn patches and rock glaciers due to the different thermal regime of blocky material (Haeberli et al., 2006). An additional source of debris is recycled material from lateral moraines. These moraines become unstable as the glacier lowers and permafrost thaws, enabling moraine material to fall onto the glaciers (Nakawo et al., 1986). Debris eroded from the glacier bed may also become entrained into the ice. The extent of subglacial entrainment depends mainly on thermal regime and substrate erodibility.

Distribution of debris

The spatial distribution of debris in and on the glacier is the result of three main processes (supraglacial, englacial, and subglacial; Hambrey et al., 2008, Nakawo et al., 1986), and depends primarily on the location of the entrainment zone and the transport of the sediment away from it. Debris supplied in the accumulation zone is buried by snow and becomes entrained into the ice, while debris supplied in the ablation zone usually remains on the surface due to emergent glacier flow. Debris supplied from below generally remains in the ice. However, some basal or even subglacial debris can also reach the glacier surface along shear horizons and become supraglacial debris (Hambrey et al., 2008). On the other hand, debris can become entrained in crevasses and then into the ice by falling from the surface (Gulley and Benn, 2007). The distribution of debris is usually inhomogeneous. Debris from point sources usually accumulates in planar shapes (debris septa) or as discrete bodies of variable geometry. Supraglacial debris can cover whole or substantial parts of the glacier tongue (debris-covered glacier), given low surface slope values (Paul et al., 2004). Supraglacial debris cover can also form at the margin of glaciers