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Research paper

Chronology of Lateglacial ice flow reorganization and deglaciation in the Gotthard Pass area, Central Swiss Alps, based on cosmogenic ^{10}Be and *in situ* ^{14}C

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ABSTRACT

We reconstruct the timing of ice flow reconfiguration and deglaciation of the Central Alpine Gotthard Pass, Switzerland, using cosmogenic ^{10}Be and *in situ* ^{14}C surface exposure dating. Combined with mapping of glacial erosional markers, exposure ages of bedrock surfaces reveal progressive glacier downwasting from the maximum LGM ice volume and a gradual reorganization of the paleoflow pattern with a southward migration of the ice divide. Exposure ages of ~16–14 ka (snow corrected) give evidence for continuous early Lateglacial ice cover and indicate that the first deglaciation was contemporaneous with the decay of the large Gschnitz glacier system. In agreement with published ages from other Alpine passes, these data support the concept of large transection glaciers that persisted in the high Alps after the breakdown of the LGM ice masses in the foreland and possibly decayed as late as the onset of the Bølling warming. A younger group of ages around ~12–13 ka records the timing of deglaciation following local glacier readvance during the Egesen stadial. Glacial erosional features and the distribution of exposure ages consistently imply that Egesen glaciers were of comparatively small volume and were following a topographically controlled paleoflow pattern. Dating of a boulder close to the pass elevation gives a minimum age of 11.1 ± 0.4 ka for final deglaciation by the end of the Younger Dryas. *In situ* ^{14}C data are overall in good agreement with the ^{10}Be ages and confirm continuous exposure throughout the Holocene. However, *in situ* ^{14}C demonstrates that partial surface shielding, e.g. by snow, has to be incorporated in the exposure age calculations and the model of deglaciation.

1. Introduction

Reconstructing the extent and dynamics of past glaciers provides valuable information on paleoclimate conditions. The extent of a glacier is determined by the mass balance, which is largely controlled by climate. Variations in the ice volume and glacial expansion therefore give evidence for paleoclimate changes (Oerlemans et al., 1998; Oerlemans, 2001). Being comparatively small ice bodies, Alpine glaciers are particularly sensitive palaeoclimate indicators that respond rapidly to changes in temperature and/or precipitation (Kerschner, 2005).

The onset of the Alpine Lateglacial is defined as the time of massive downwasting of the large valley glaciers built up during the Last Glacial Maximum (LGM; late Würmian) (Penck and Brückner, 1901/1909; Reitner, 2007). Deglaciation of the Alps is thought to have occurred rapidly with retreat of the foreland piedmont glaciers closely followed by the disintegration of the Central Alpine ice cap (Florineth and Schlüchter, 1998; Schlüchter, 2004). Although climate was warming gradually, short climatic fluctuations and cold phases repeatedly interrupted the general trend of ice decay (e.g. Reitner, 2007; Schmidt et al., 2011). These fluctuations are associated with several Lateglacial stadials determined by glacier stillstands and readvances before the beginning of the Holocene warming (Ivy-Ochs et al., 2008 and references therein).

Classically, former extents of Alpine glaciers and related climate changes have been dated by radiocarbon from organic material, mainly retrieved from glacial deposits in the Alpine foreland (e.g.

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Heuberger, 1966; Patzelt, 1972; Fliri, 1973; Draxler, 1977; Bortenschlager, 1984; Geyh and Schreiner, 1984; Kerschner, 1986; Hajdas et al., 1993; Schlüchter and Röthlisberger, 1995; van Husen, 2004; Keller and Krayss, 2005; Reitner, 2007 and references therein). These data are synchronized with relative chronologies constructed from morpho- and lithostratigraphic observations and fossil and pollen evidence (e.g. Penck and Brückner, 1901/1909; Keller and Krayss, 1987; Ammann et al., 1994; Wohlfarth et al., 1994; van Husen, 2000; Vescovi et al., 2007; Preusser et al., 2011; van Husen and Reitner, 2011). In the last two decades, direct dating of glacial surfaces and moraine deposits by cosmogenic nuclides has provided further valuable information on the chronology of past glacial extents and the timing of glacier retreat (Ivy-Ochs et al., 1996, 2006b, 2007, 2008; Kelly et al., 2006; Hormes et al., 2008; Böhlert et al., 2011; Federici et al., 2012; Reuther et al., 2011). From these data a detailed chronology for the early recession of the LGM ice and the Lateglacial readvances ('stadials') in the Alpine foreland and the inner Alpine valleys was established.

In the high Alps, however, the timing of the breakdown of the LGM ice cap and the extent of Lateglacial local ice is less well constrained. From mapping of glacial trimlines and further glacial erosional features the presence of large, precipitation-controlled ice domes during the LGM has been proposed (Florineth and Schlüchter, 1998, 2000; Kelly et al., 2004a). From the inner Alpine accumulation areas, ice streams expanded into the valleys and glacier transfluences developed at several high Alpine passes (Florineth, 1998; Florineth and Schlüchter, 1998). It is assumed that this ice flow configuration persisted until the piedmont glaciers in the Alpine forelands retreated from their maximum extent (Florineth and Schlüchter, 1998; Kelly et al., 2006). Thus, major downwasting of the inner Alpine ice masses and the termination of glacier transfluence over the high passes were thought to be roughly synchronous to the onset of ice-free conditions in the foreland valleys. However, there is evidence that the high Alps did not deglaciate by that time but that ice persisted locally at high elevations until the beginning of the Bølling/Allerød interstadial (~15–14 ka BP) or even longer (Kelly et al., 2006; Böhlert et al., 2011).

To improve the knowledge about the Lateglacial ice decay in the high Alps and better understand the effect of climate changes on the high Alpine mountain glaciers and ice caps, we performed surface exposure dating of glacially modified bedrock on the Gotthard Pass, central Swiss Alps, using cosmogenic ^{10}Be and *in situ* ^{14}C. Combining ^{10}Be dating with *in situ* ^{14}C analyses takes advantage of the short half-life of ^{14}C (5730 years) which allows to recognize episodes of surface burial as well as constant surface shielding. *In situ* ^{14}C can therefore provide information on the extent of Holocene ice on the Gotthard Pass and can be used to evaluate the necessity of snow shielding corrections for exposure dating in an Alpine environment.

To reconstruct the paleoflow pattern, we combine surface exposure dating with detailed mapping of glacial erosional features in the Gotthard pass area. This yields a detailed chronology of the LGM ice surface lowering and local glacier readvances in conjunction with a progressive reorganization of the glacial ice flow pattern from the LGM until the Holocene.

2. Chronology of the Lateglacial ice decay in the Alps

By ~21 ka the Alpine foreland piedmont glaciers had started to retreat from their maximum position (Schlüchter, 1988, references therein; Schlüchter, 2004; Ivy-Ochs et al., 2004; Preusser, 2004; Keller and Krayss, 2005; Preusser et al., 2011). Deglaciation of the foreland and the inner Alpine valleys by ~19–18 ka marks the beginning of the Alpine Lateglacial (Lister, 1988; Wessels, 1998; van Husen, 2004; Reitner, 2007). According to the work of Reitner (2007) there is no evidence for a climate-driven glacier readvance during the early Lateglacial and therefore the concept of early Lateglacial stadials ('Bühl' and 'Steinach') should be abandoned and replaced by the expression 'phase of early Lateglacial ice decay'. An early Lateglacial phase of warming at ~18.0–17.5 cal ka BP was recorded regionally from southern Alpine macrofossil and pollen data (cf. Vescovi et al., 2007), and by the δ^{18}O in Greenland ice cores (Björk et al., 1998). In the following, a series of prominent moraines were deposited throughout the Lateglacial until the beginning of the Holocene within the Central and Eastern Alps (Maisch, 1981, 1982; Kerschner and Berktold, 1982; Kerschner, 1986).

The first pronounced glacier readvance occurred during the Gschnitz stadial. A maximum age for Gschnitz advances is given by a radiocarbon age of 15.4 ± 0.5 ^{14}C ka BP (19.6–17.6 cal ka BP) from the Eastern Alps (Draxler, 1977). Consistently, exposure ages obtained by ^{10}Be dating of moraine boulders at the type locality in Trins (Gschnitz Valley, Tyol, Austria) as well as from a Gschnitz terminal moraine deposit from the Maritime Alps (Italy) have provided mean ages around 17 ka (Ivy-Ochs et al., 2006a; Federici et al., 2012). Note that these ages and all published ^{10}Be exposure ages discussed below have been renormalized to the 07KNSTD standardization and recalculated using a ^{10}Be spallogenic production rate of 3.93 at g^{-1} y^{-1} (see Section 4.3), without including corrections for erosion or snow, even if these were applied for the original ages. The Gschnitz advance was followed by pronounced glacier downwasting before smaller readvances of the Clavadel/Senders and Daun stadials (Maisch et al., 1999; Ivy-Ochs et al., 2006b; Kerschner, 2009). These readvances were followed by marked ice decay during the Bølling/Allerød interstadial (~14.7–12.9 ka; Björk et al., 1998; Vescovi et al., 2007; Ivy-Ochs et al., 2008) which ended abruptly with the beginning of the Younger Dryas cold period (Ammann et al., 1994; references therein; Rasmussen et al., 2006). Widespread glacier readvance during the Younger Dryas is documented by series of Egesen stadial moraine complexes deposited throughout the Alpine valleys (Kerschner et al., 2000 and references therein). Several surface exposure ages have been obtained from different moraine complexes in Switzerland and Italy. ^{10}Be ages range between ~13.9 and 10.6 ka and are interpreted to give the timing of moraine stabilization during various glacier readvances of the Egesen stadial (Ivy-Ochs et al., 1996, 1999, 2006b; Kelly et al., 2004b, 2006; Federici et al., 2007; Hormes et al., 2008). Several radiocarbon ages of ~11 cal ka BP (summarized and calibrated in Ivy-Ochs et al., 2008) are consistent with ^{10}Be ages and trace the final downwasting of Egesen glaciers. Continued glacier activity into the earliest Holocene is suggested from moraine deposition and rock glacier activity (Fraedrich, 1979; Frauenfelder et al., 2001; Ivy-Ochs et al., 2006b, 2009; references therein; Kerschner and Ivy-Ochs, 2008).

3. Study area

The Gotthard pass is located in Central Switzerland with a pass elevation of 2106 m a.s.l. (Fig. 1). Since the historic past, it has been one of the most important routes traversing the Alps in north-south direction. The Gotthard region comprises crystalline rocks of the Gotthard Massif that forms a part of the crystalline basement of the Swiss Alps. Granitic rocks in the study area are associated with the Fibbia granite gneiss, which is separated by paragneisses from the Gamsboden granite gneiss further north (Labhart, 2009). The Fibbia granite gneiss exhibits a porphyritic texture and syenogranitic composition (Sergeev et al., 1995; Debon and Lemmet, 1999). Plutonic rocks of the Gotthard massif are of Variscan age and were deformed during Alpine orogeny at greenschist to amphibolite facies conditions (Frey et al., 1980; Labhart, 1999).

The pass area shows a typical U-shaped cross section and is characterized by highly polished granitic surfaces and abundant glacial erosional features (Fig. 2). Trimlines at 2640 m a.s.l. on the