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Autor(en): **Ivy-Ochs, Susan / Schlüchter, Christian / Kubik, Peter W.**

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The exposure age of an Egesen moraine at Julier Pass, Switzerland, measured with the cosmogenic radionuclides ^{10}Be, ^{26}Al and ^{36}Cl

SUSAN IVY-OCHS[1], CHRISTIAN SCHLÜCHTER[2], PETER W. KUBIK[3], HANS-ARNO SYNAL[3], JÜRG BEER[4] & HANNS KERSCHNER[5]

Key words: Younger Dryas, Egesen moraines, Julier Pass (Graubünden), exposure dating, cosmogenic isotopes, ^{10}Be, ^{26}Al, ^{36}Cl

ABSTRACT

The Egesen moraine complex on the east side of Julier Pass has a mean exposure age of 11,100 years. The exposure age is based on accelerator-mass-spectrometry-measured concentrations of ^{10}Be, ^{26}Al and ^{36}Cl in six different boulders. Three of the boulders lie along the sharp-crested outer lateral moraine and have a mean exposure age of 11,800 years. This moraine wall represents the glacier's response to the climatic collapse at the very beginning of the Younger Dryas. The other three boulders are located along a hummocky boulder band that cross-cuts the outer moraine along its terminal front. The exposure age of the inner moraine is 10,400 years.

ZUSAMMENFASSUNG

Der Egesenmoränenkomplex östlich des Julierpasses besitzt ein mittleres Expositionsalter von 11,100 Jahren. Das Expositionsalter basiert auf mittels Beschleunigermassenspektrometrie bestimmten Konzentrationen an ^{10}Be, ^{26}Al and ^{36}Cl in sechs verschiedenen Blöcken. Drei dieser Blöcke befinden sich auf der steilgeböschten äusseren Seitenmoräne und zeigen ein mittleres Expositionsalter von 11,800 Jahren. Dieser Moränenwall wurde durch den Gletscher zur Zeit des Klimakollapses zu Beginn der Jüngeren Dryas gebildet. Die drei anderen Blöcke befinden sich auf einem höckrigen Blockstreifen, der die äussere Moräne entlang ihrer Stirnseite schneidet. Das Expositionsalter dieser inneren Moräne beträgt 10,400 Jahre.

Introduction

During the Last Glacial Maximum, coalescing piedmont glaciers, e.g. the Rhine and Rhone glaciers, almost completely inundated the northern foreland of the Swiss Alps with ice (Jäckli 1970). Following attainment of their maximum extent at around 20,000 years ago, glacial ice vanished rapidly from the foreland in only a few thousand years (Schlüchter 1988). Radiocarbon dated sediments from foreland lakes indicate completely ice-free conditions no later than 14,600 at Zurichsee (Fig. 1) (Lister 1988) and 14,200 at Soppensee (Hajdas et al. 1993) radiocarbon years. The Oldest Dryas vegetation zone

[1] Institut für Teilchenphysik; Ingenieurgeologie; ETH Hönggerberg, CH–8093 Zürich
[2] Geologisches Institut, Universität Bern, CH–3012 Bern
[3] Paul Scherrer Institut c/o Institut für Teilchenphysik, ETH Hönggerberg, CH–8093 Zürich
[4] EAWAG, CH–8600 Dübendorf
[5] Institut für Geographie, Innrain 52, Universität Innsbruck, A–6020 Innsbruck

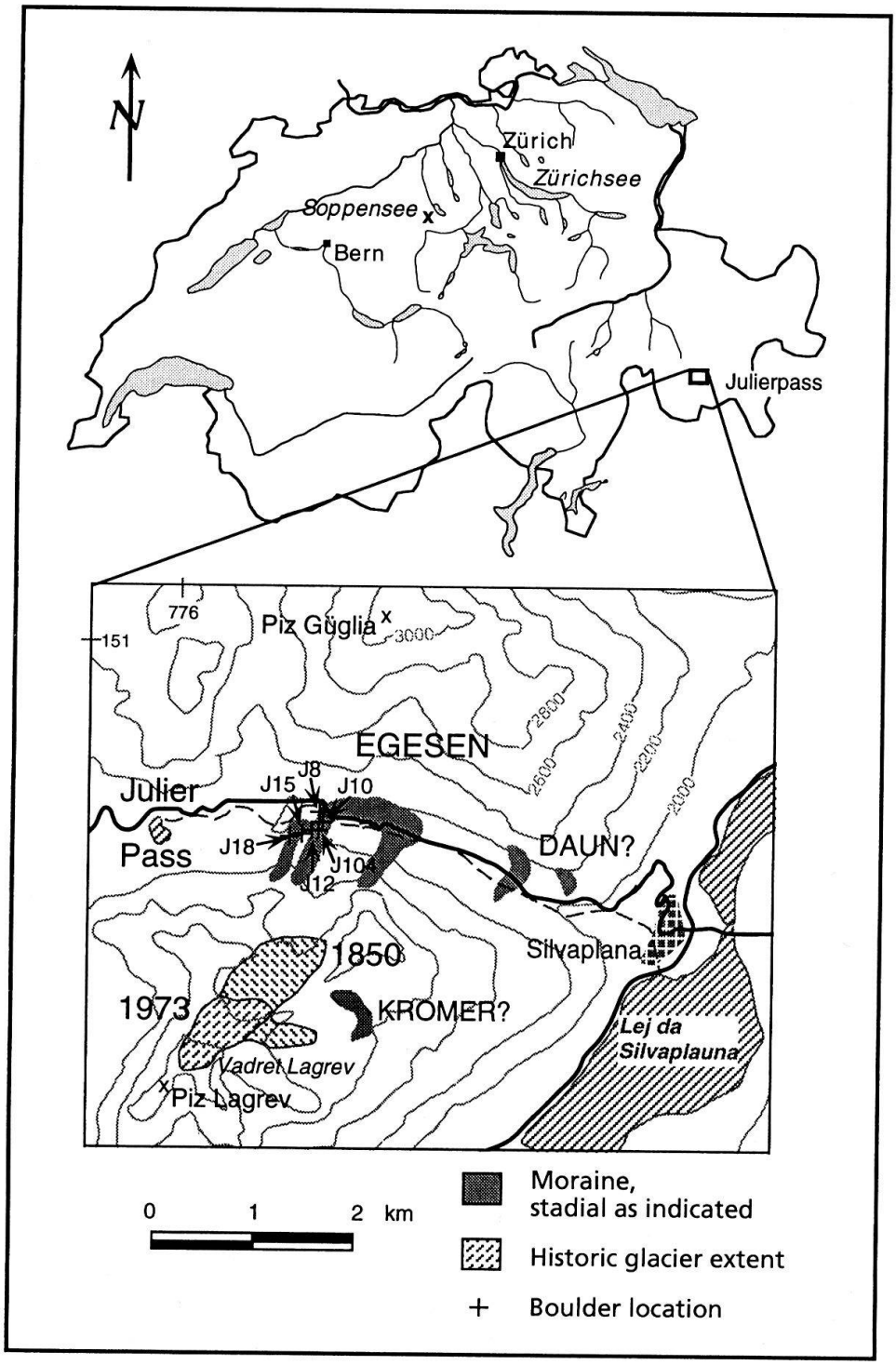

Fig. 1. Map of Switzerland showing the location of Julier Pass and some of the sites discussed in the text. Schematic map of the Julier Pass region with the historic glacier extents and the moraines formed during Late Glacial glacier expansions (modified from Cornelius 1935/1950/1951, Suter 1981, Maisch 1995). The outer lateral moraine is the western band in the Egesen complex. Sampled boulder locations are shown as crosses.

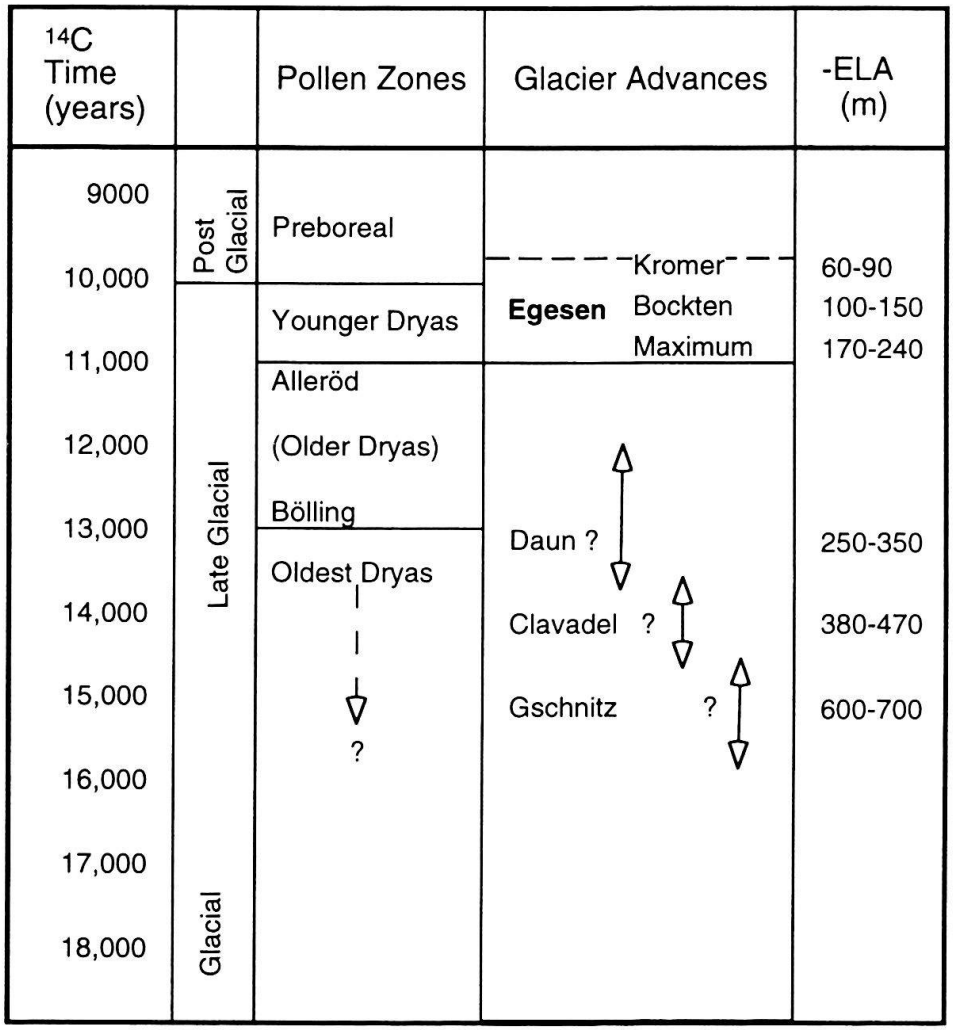

Fig. 2. Summary of the Late Glacial vegetation zones (Mangerud et al. 1974), glacier re-advances and the corresponding ELA depressions (Maisch 1982).

(Fig. 2) marks the influx of pioneer tundra-type vegetation (e.g. Burga 1988). Oldest Dryas sites, with minimum dates on the order of 14,000 to 13,000 radiocarbon years (Furrer et al. 1987 and references therein), located in valleys deep within the Alps, indicate absence of glacial ice in many alpine valleys by that time. The Oldest Dryas was followed by a period of time warm enough for pine trees, the Bølling/Allerød, which was cut short by the cold event known as the Younger Dryas.

The extremely rapid decay of the Alpine glaciers was punctuated by minor readvances which were very much confined to the Alps proper. These stadia are classified by the moraines which were formed, for example Bühl (Penck & Brückner 1901/1909), Steinach (Senarclens de Grancy 1958), Gschnitz (Penck & Brückner 1901/1909), Clavadel (Maisch 1981), Daun (Penck & Brückner 1901/1909, Heuberger 1966, Mayr & Heuberger 1968) and Egesen (Kinzl 1929, Heuberger 1966) (Fig. 2). The moraine sequence is fairly consistent from glacier to glacier and has been extensively mapped in both the Swiss and Aus-

trian Alps (e.g. Heuberger 1966, Mayr & Heuberger 1968, Kerschner 1978, Maisch 1981, 1987, Suter 1981, Müller et al. 1981, Müller 1984). General correlation from valley to valley has been accomplished with carefully established geomorphologic characteristics combined with the inferred drops in equilibrium line altitude (ELA) from the 1850 reference level (Gross et al. 1978, Maisch 1992).

The Egesen moraines are probably the most marked and easily identifiable of the series. They were first described by Kinzl (1929) in the Stubai Mountains to the southwest of Innsbruck (Austria). There, the type locality is at the Egesengrat and its surroundings near the Dresdner Hütte. Originally, Kinzl did not consider the Egesen stade to have been an important advance. Later, in 1966, Heuberger showed that Egesen moraines are widely distributed and were formed during a prominent glacier advance, which was clearly separated from the earlier Daun stade. During the Bølling/Allerød milder climate period, the glaciers probably retreated back to roughly the 1850 level (Maisch 1982). In the central Alps Egesen moraines are classified by their location at the ca. 200 m ELA difference (Patzelt 1972, Gross et al. 1978, Maisch 1987) and by their fresh and frequently blocky form in comparison to the more subdued older Daun moraines (Heuberger 1966).

Geomorphologically, Egesen moraines clearly represent the last cold event (i.e. Younger Dryas) prior to the many advances of Postglacial order of magnitude, which characteristically reached only the "Little Ice Age" – extent (Patzelt & Bortenschlager 1973, Holzhauser 1984, 1987). Egesen moraines do not show any evidence of permafrost or solifluction overprint but retain their fresh and blocky morphology, depending on lithology. This is in clear contrast to the older moraines at similar altitudes, particularly the Daun moraines (Heuberger 1966) which were modified by periglacial processes following deposition as a result of the very cold conditions during the Younger Dryas.

Egesen moraines are characterized by several nested and slightly cross-cutting moraines (Heuberger 1966, Kerschner 1978, Maisch 1981, 1987). In many places three, sometimes four distinct series of moraines can be found. They indicate that the climatic deterioration was at least three phased. Basis radiocarbon dates from bog deposits in the tongue areas of several Egesen age moraines in the Austrian Alps are in the early Preboreal time range (Patzelt 1972, Kerschner 1978, Bortenschlager 1984). In the Val de Nendaz, the last phase of the Younger Dryas could be found by pollen analysis in lake deposits within the moraines of the second phase of the Egesen stade (Küttel 1979, Müller et al. 1981). Egesen moraines thus correspond most probably to the Younger Dryas. There are no radiocarbon dates from the Vadret Lagrev Egesen moraine at Julier Pass.

For the older moraines, organic material is lacking and ^{14}C dating has not been possible. Thus no precise time period can be assigned to any of these moraines except the Egesen. The earlier stadia can only be placed in time somewhere between the Last Glacial Maximum and the Younger Dryas.

Exposure dating

Sample preparation

Cosmic rays encountering an exposed rock surface induce nuclear reactions within the mineral lattices yielding "cosmogenic isotopes", e.g. ^{10}Be ($t_{1/2}$=1,500,000 years), ^{26}Al ($t_{1/2}$=716,000 years) and ^{36}Cl ($t_{1/2}$=301,000 years) (e.g. Lal 1991, Cerling & Craig 1994,

Kurz & Brook 1994, Zreda & Phillips 1994). The concentration of the cosmogenic isotope, which is measured by accelerator mass spectrometry (AMS), is a direct measure of the length of time the geologic surface has been exposed. Equation 1 describes the growth of a radionuclide concentration with time in the mineral grains of the exposed rock surface:

$$N = \frac{P}{\lambda + \frac{\rho\varepsilon}{\Lambda}}\left(1 - e^{-(\lambda+\rho\varepsilon/\Lambda)T}\right) \quad (1)$$

where N is the number of atoms/gram, P is the local production rate in atoms/g.yr, T is the length of time the surface has been exposed in yr, λ is the decay constant of the radionuclide in yr^{-1}, ρ the rock density in g/cm^3 (2.7 for granite and granodiorite), ε, the erosion rate in cm/yr, and Λ is the cosmic ray attenuation length in g/cm^2 (150 g/cm^2; Brown et al. 1992).

Quartz is well suited to ^{10}Be and ^{26}Al exposure dating studies because it has a low inherent Al content allowing measurement of $^{26}Al/^{27}Al$. Its tight crystal structure makes it a closed system to either loss or gain (other than in situ production) of the isotope of interest (Kohl & Nishiizumi 1992).

We sampled the top surfaces of the largest boulders on the crests of the moraines (more than 1 meter higher than the surrounding sediment) by using a hammer and a chisel. Boulders with obvious signs of spalling or rolling over were avoided. Altitudes were determined from the topographic map and with an altimeter. The boulders were predominantly petrologic facies of the Julier Granite (Bühler 1983) including granodiorite (J8, J10, J104, J18), aplite (J12), and diorite (J15). Thin section examination indicated an average quartz grain size of 0.5–0.7 mm. The rocks are fairly fresh, with the highest degree of weathering in the feldspars (saussuritization), yielding the greenish cast typical of the Julier Granite.

Ca. 0.5–1.0 kg of each rock sample were crushed and sieved. The <0.4 mm size fraction was used for ^{36}Cl sample preparation from whole rock, while the 0.4–1.00 mm size fraction was used for the separation of quartz to be used for ^{10}Be and ^{26}Al. A pure quartz fraction is gained by selective chemical dissolution (Kohl & Nishiizumi 1992). With this method the weak HF solution dissolves the feldspars and micas faster than the quartz, purifying the quartz fraction. We use at least five HF steps so the quartz fraction is very pure, reflected as well by the low Al concentration (e.g. 79 ppm in J18). The elegance of this method is that any contaminating meteoric ^{10}Be from the atmosphere, which can be several orders of magnitude greater than the in situ component, is etched away from the surfaces of the quartz grains (Kohl & Nishiizumi 1992). ^{9}Be carrier was added to the dried and weighed pure quartz (40–70 grams for each sample), which was then completely dissolved with concentrated HF. Be and Al were separated using cation exchange. The hydroxides were precipitated then baked to the oxides, Al_2O_3 and BeO, at 850 °C. $^{10}Be/^{9}Be$ and $^{26}Al/^{27}Al$ were measured by AMS at the ETH/PSI tandem facility in Zürich. Stable Al was measured by ICP-AES (inductively-coupled plasma atomic emission spectroscopy) using three standard additions for each sample.

The <0.4 mm size fraction of the crushed whole rock was used for ^{36}Cl sample preparation. It is important to note that the exact same crushed rock sample was used for all three isotopes. Sample preparation for ^{36}Cl followed the method of Zreda (1994). To leach any possible meteoric ^{36}Cl, the <0.4 mm fraction was left overnight in a solution of weak HNO_3, then rinsed at least five times with 18 MΩ water. On the order of 70 g of dry whole rock were placed in teflon bottles with concentrated HNO_3 and HF. The bottles were kept in a 90 °C water bath until the rock had dissolved completely, which usually took several days. The contents of the bottle were centrifuged cold (10 °C) in teflon centrifuge bottles in order to separate reprecipitated silica gel from the dissolved rock found in the HF solution. The liquid was decanted into teflon centrifuge tubes, then $AgNO_3$ was added to the HF solution to precipitate AgCl. The solution was left at least 24 hours to allow nucleation of the AgCl. After centrifuging and decanting, the AgCl was redissolved with NH_4OH and unwanted hydroxides were precipitated, then separated with centrifuging. $Ba(NO_3)_2$ was added to the solution to remove sulfur by formation of $BaSO_4$ (^{36}S is an interfering isobar in AMS measurements) and left again at least 24 hours. The $BaSO_4$ was removed by filtering, then the AgCl was reprecipitated, dried and pressed into tantalum target holders. $^{36}Cl/Cl$ ratios were measured by AMS at Zürich. Major elements, Gd, B and Cl were measured using XRF, prompt gamma neutron activation and ion selective electrode, respectively (Tab. 1). These elements were all measured on the rinsed sample material.

Element	J10	J12	J15
weight %			
SiO_2	73.5	77.5	40.4
Al_2O_3	11.3	11.5	18.1
Fe_2O_3	3.91	1.52	15.6
MnO	0.07	0.02	0.16
MgO	1.17	0.43	6.6
CaO	1.85	0.94	10.2
Na_2O	1.94	2.99	1.71
K_2O	3.9	4.04	0.72
TiO_2	0.38	0.12	2.09
P_2O_5	0.1	0	0.06
total	98.12	99.06	95.64
ppm			
B	13.5	10	13
Gd	5.5	2	4.5
Cl	189	115	213
U	2.8		
Th	6.7		

Tab. 1. Major and minor element concentrations for the rocks analyzed for ^{36}Cl.

Exposure age Calculations

Ages are calculated by solving Eq. 1 for time (T), and assuming no erosion of the rock surface (ε=0):

$$T = \frac{ln\left(1 - \frac{N\lambda}{P}\right)}{-\lambda} \quad (2)$$

We have used the production rate of 6.0 ^{10}Be and 36.7 ^{26}Al atoms per gram SiO_2 per year at high latitude (>60°) and sea level for the age calculations (Nishiizumi et al. 1989). The uncertainty of the ^{10}Be and ^{26}Al production rates have been estimated at less than 10% (Masarik & Reedy 1995) and 5–7% (Lal 1991).

For ^{36}Cl there are three main production mechanisms, i.e. spallation of Ca and K to form ^{36}Cl and neutron capture by ^{35}Cl to form ^{36}Cl (e.g. Zreda & Phillips 1994, Zreda 1994). Therefore, the production rate must be calculated individually for each rock composition. We have used the elemental production rates determined by Phillips et al. (1996). They are 1280 ^{36}Cl atoms per kg rock per wt. % K_20 per year and 530 ^{36}Cl atoms per kg rock per wt. % Ca0 per year (high latitude and sea level). The errors have been estimated at 9% (Zreda 1994).

In order to calculate the production of ^{36}Cl due to neutron capture by ^{35}Cl one must first calculate the fraction of neutrons absorbed by the rest of the rock. We have calculated this fraction using the measured major element concentrations as well as the concentrations of B and Gd (Tab. 1) which are significant neutron absorbers and the cross section data from Zreda (1994). The fraction available for production of ^{36}Cl from Cl is then multiplied by the neutron flux, 313,500 neutrons per kg rock per year (Zreda 1994) and the measured Cl concentration for that rock. The resulting ^{36}Cl production rates for each rock range from ca. 11 to 14 ^{36}Cl atoms/gram rock per year. We have corrected for subsurface (non-cosmogenic) production (Fabryka-Martin 1988) of ^{36}Cl based on the measured U and Th concentrations in J10. In all cases, this amounts to a correction of less than 2 %.

The production rate for each of the three isotopes must then be scaled to the latitude (Julier Pass 46°30') and altitude of the samples (Lal 1991). The uncertainty of the scaling is thought to be less than 10% (Lal 1991, Masarik & Reedy 1995).

We have made two additional corrections, one for shielding and one for the thickness of the sample analyzed (Tab. 2). The production of cosmogenic isotopes is dependent on the cosmic rays that the rock surface actually sees. Therefore, reduction of the cosmic rays by ridges and mountain peaks directly surrounding the sampling site, reduces the production rate accordingly. To the west and east from Julier Pass the shielding was zero, while the ridges related to Piz Güglia to the north of the moraine and Piz Lagrev to the south (Fig. 1) shielded the horizon by roughly 25 degrees. The precise angle to the horizon and the number of degrees of the quadrant which were affected were then used to calculate (Nishiizumi et al. 1989) the production rate reduction. This correction was only 3.5 % for all samples. For boulders whose original surface was dipping, the shielding cor-

Tab. 2. AMS-mesuared concentrations of ^{10}Be, ^{26}Al and 36Cl and calculated exposures ages.

Boulder No.	Lithology	Thickness (cm)	Altitude (m)	Isotope	Atoms/gram	Age (years)	Mean boulder age	Moraine age
J18	granodiorite	4	2210	^{10}Be	3.50×10^5 *	11,650 ± 910	11,800 ± 840	Outer moraine:
				^{26}Al	2.29×10^6 *	12,640 ± 2150		11,820 ± 500
J15	diorite	2	2195	^{36}Cl	6.84×10^5	11,490 ± 1150	11,490 ± 1150	
J12	aplite	4	2200	^{10}Be	3.60×10^5 *	12,060 ± 1550	12,210 ± 880	
				^{36}Cl	7.72×10^5	12,270 ± 1060		
J104	granodiorite	12	2185	^{10}Be	2.86×10^5 *	10,390 ± 760	10,390 ± 760	Inner moraine
J10	granodiorite	8	2185	^{10}Be	2.87×10^5	10,050 ± 770	10,480 ± 620	10,380 ± 400
				^{36}Cl	8.67×10^5	11,260 ± 1030		
J8	granodiorite	6	2175	^{10}Be	2.96×10^5	10,260 ± 700	10,260 ± 700	

* data from Ivy-Ochs et al. 1995.

rections which includes the dip angle as well, were also calculated. This only affects J15 (32° dip to the southeast) making the total reduction of the production rate due to shielding 5.2 % for J15.

The atoms which are produced in a rock surface falls off exponentially with depth due to the attenuation of cosmic rays penetrating matter. The production rate for each sample must be corrected based on the thickness of the piece of rock which was crushed and dissolved (Tab. 2). For ^{36}Cl production, spallation components (production of ^{36}Cl from Ca and K) have been corrected for separately than neutron capture (based on Fabryka-Martin 1988 and Zreda 1994).

No correction has been made for erosion of the surfaces, estimated to be small based on geomorphologic appearance, for example lack of evidence of spalling. On the other hand, a fine-grained till matrix may have covered the boulders right after moraine formation. We hoped to avoid this problem by sampling boulders on the crests of the moraines. No correction has been made for intermittent snow cover. The position of the boulders on top of the moraine allows much of the snow that falls to be blown off. In recent years winter snow cover on these boulders has been nil.

The exposure age of the Julier Pass Egesen moraine complex

We have determined nine exposure ages from six different boulders located on the crest of the Egesen moraine complex at Julier Pass (Tab. 2). The errors listed are the 1σ measurement errors. They include both the statistical (counting) error and the error due to the normalization to the standards and blanks. A 5% sample reproducibility error has been included in the error on each exposure age. This is based on the reprocessing of several different rock samples (see Ivy-Ochs 1996 for details). The ^{36}Cl age includes the error from the total Cl measurement (8%) and the ^{26}Al age the total Al measurement (1%).

Fig. 3. Air photo of the terminal zone of the Vadret Lagrev glacier (east side of Julier Pass) during the Egesen stade. The outer moraine is found in the center foreground (arrow) (photo: C. Schlüchter).

Exposure ages have been determined for sample J18 with both ^{10}Be and ^{26}Al from a quartz mineral separate. The two ages agree quite well within the given errors. We have only measured ^{36}Cl in sample J15 because it contained very little quartz. Samples J10 and J12 have both ^{10}Be and ^{36}Cl exposure ages which also show good agreement. The agreement is noteworthy since ^{36}Cl was extracted from a whole rock sample, while ^{10}Be was measured in quartz. Additionally, the published production rates for these three isotopes have been determined quite differently. Production rates for ^{10}Be (and ^{26}Al) were determined in a glacially polished surface where the calendar age of the surface was inferred based on regional correlations (Nishiizumi et al. 1989). While the ^{36}Cl production rates were calculated based on a suite of rock samples from several different sites and with ages ranging from 2,000 to 50,000 years (Phillips et al. 1996). For ^{10}Be and ^{26}Al we have used the presently accepted production rates (Nishiizumi et al. 1989) but recognize that reevaluation of these production rates may be called for (Clark et al. 1995). Based on our Julier Pass data, one might conclude that in this time range, the published produc-

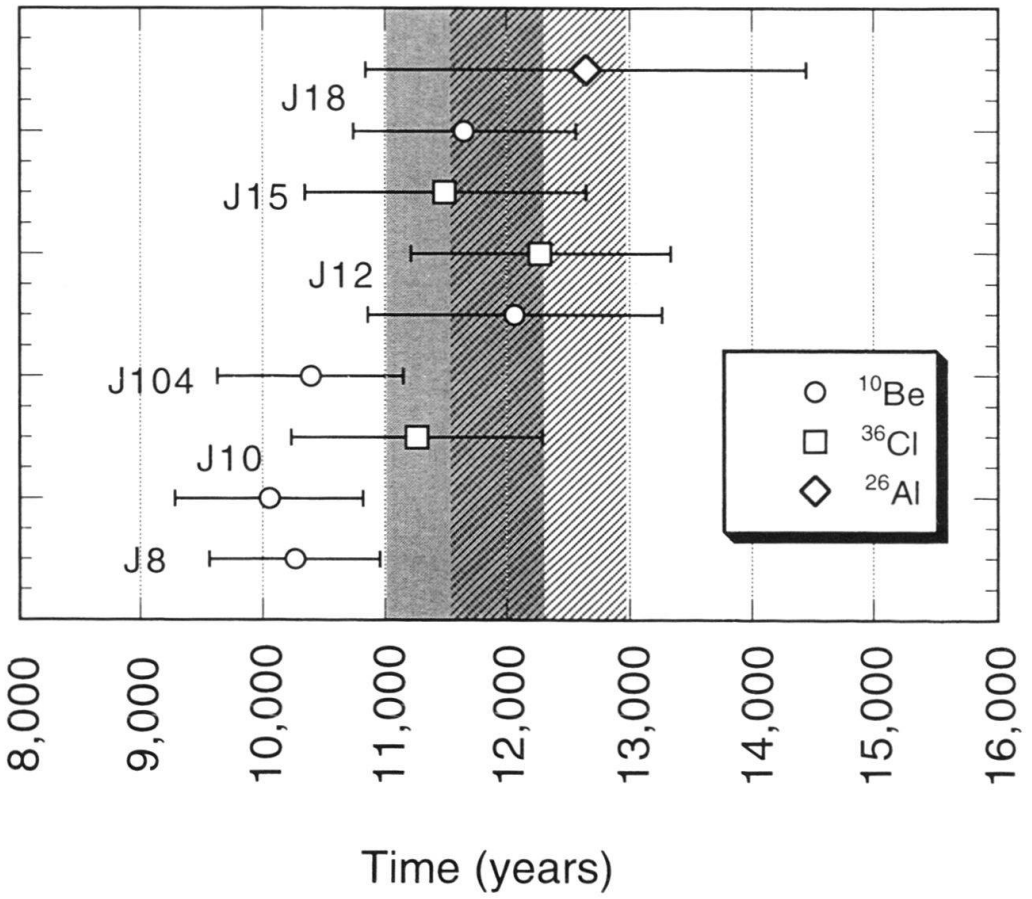

Fig. 4. Plot of ^{10}Be, ^{26}Al and ^{36}Cl exposure ages determined for six boulders from the Egesen moraine complex at Julier Pass. The shaded band represents the Younger Dryas as determined by varve counting in sediment cores from Soppensee (Hajdas et al. 1993); the cross-hatched area shows the Younger Dryas as recorded in ice cores (Alley et al. 1993, Johnsen et al. 1992).

tion rates for all three isotopes and the latitude/altitude scaling factors are probably less than 10 % off.

The exposure ages we have measured range from 10,050 to 12,640 years. When more than one isotope was determined from a single boulder, its exposure age was obtained by calculating the weighted mean (weighted based on the respective errors from each measurement). The overall mean age of the moraine complex is 11,100 ± 700 years.

The terminal zone of the Egesen moraine at Julier Pass consists of several cross-cutting boulder bands and a tongue area filled with blocks, all apparently having been deposited relatively closely spaced in time. The outer lateral moraines are the highest and sharpest features. Their well-defined shape and considerable size indicates an abrupt glacier advance due to rapid temperature drop where the glacier remained in its maximum position for a rather long period of time. Boulders J18, J15, J12 lie along the crest of the sharp outer lateral moraine. This moraine hooks around to the east and is cross-cut by a later moraine made up of an ill-defined band of boulders (Fig. 3). Samples J8, J10 and J104 are from this inner moraine. The exposure ages from the three rocks from the outer

moraine give an average of 11,800 ± 500. For the three rocks from the inner moraine 10,400 ± 400 is obtained. The exposure ages are in line with the field data, as they indicate that the inner moraine is slightly younger than the outer moraine.

The beginning of the Younger Dryas was marked by an extreme cold snap (for example seen in $\delta^{18}0$ in GRIP, Dansgaard et al. 1993). It may be that this correlates to the sharp outer moraine wall. The later inner moraine can actually be interpreted as a transitional form between a moraine and an ice-rich rock glacier, which, under permafrost conditions, might also have contained some remnant glacier ice. In any case, it is a typical feature of ice-rich permafrost. These landforms are characteristic for the later phases of the Egesen stade at small glaciers in the continental parts of the central Alps. They suggest that climatic conditions remained cold and became successively drier during the Younger Dryas. As a result, glaciers were starved and wide-spread permafrost features developed (Kerschner 1982, 1985). Detailed field mapping combined with further exposure dating will shed more light on these complex relationships.

As these boulders are most likely supra-glacial material originating from rockfalls onto the glacier surface, prior exposure is a definite possibility. This would happen when the rock has been exposed to cosmic rays either along the valley wall or cliff. Because of the coherence of the six boulder ages, we rule out prior exposure for the surfaces we sampled. A rock that had been exposed earlier would stick out as an older outliner.

Comparison with other Younger Dryas records

Our nine exposure ages, along with a band showing the approximate calendar age of the Younger Dryas, are presented graphically in figure 4. Calendar years in ice cores are based on layer counting (e.g. annual changes in $\delta^{18}O$ and dust) (Alley et al. 1993, Johnsen et al. 1992). At present the calendar boundaries of the Younger Dryas are slightly shifted to the older in ice cores as compared to several terrestrial records (Hajdas et al. 1995) (for example Soppensee, Switzerland, Hajdas et al. 1993, Lake Van, Turkey, Landmann et al. 1996). In Switzerland, the beginning and end of the Younger Dryas have been very precisely determined by AMS ^{14}C dating of macrofossils from the small lake Soppensee (Hajdas et al. 1993). Using the varve time scale as calendar time, the Younger Dryas as defined by the pollen signal, lasted from roughly 12,100 to 11,000 calendar years ago. This is shown by the shaded band in figure 4. The cross-hatched band is for the Younger Dryas in ice cores. Given the uncertainties in the dating for each of these records, our moraine age of 11,100 years corresponds quite well. Indeed, we are comparing very different signals (Broecker 1992) which may have different response times, i.e. pollen changes (in Soppensee), temperature change reflected by $\delta^{18}O$ shifts (in the ice cores) and a glacier's advance as recorded by a moraine, but they are all the result of the same marked drop in temperature.

Also contained in Table 3 are the ages of other moraines interpreted to correspond to the Younger Dryas which have been dated using cosmogenic isotopes. The Llyn Idwal moraine (Scotland) has been classified as Younger Dryas based on radiocarbon dating (Phillips et al. 1994). We can see that the ^{36}Cl exposure ages for both the Llyn Idwal moraine (Phillips et al. 1994, 1996) and our Julier Pass Egesen moraine agree quite well, even though in each case the data set is small. Striking agreement is as well obtained for the ^{10}Be exposure age for the Inner Titcomb Lakes moraine in the Wind River Moun-

Tab. 3. Comparison with several other Younger Dryas records.

Site	Signal	Beginning	End	Dating method	Reference
Soppensee, Switzerland	pollen change	ca.10,700-11,000	ca. 9900-10,200	^{14}C	Hajdas et al. 1993
Soppensee, Switzerland	pollen change	~12,125	~11,000	varve counting	Hajdas et al. 1993
GISP 2	$\delta^{18}O$ shift	12,940 ± 550	11,640 ± 250	layer counting	Alley et al. 1993
GRIP	$\delta^{18}O$ shift	12,700 ± 100	11,550 ± 70		Johnsen et al. 1992
Llyn Idwal, Scotland	glacier advance moraine(s) formed	11,600 ± 1300 12,900 ± 2000		^{36}Cl #	Phillips et al. 1994
Titcomb Lakes, Wyoming, U.S.A.	glacier advance moraine(s) formed	11,000 ± 770		^{10}Be *	Gosse et al. 1995
Julier Pass, Switzerland	glacier advance moraine(s) formed	11,100 ± 700 (11,800 ± 500) (10,400 ± 400)		^{10}Be, ^{26}Al, ^{36}Cl	this work

#two rocks analyzed.
*ten rocks analyzed.

tains (Wyoming, U.S.A., Gosse et al. 1995) and our Julier Pass Egesen moraine. Since in both cases the same production rates (Nishiizumi et al. 1989) and the same scaling factors (Lal 1991) were used, the two geographically widely-spaced events can be said to be contemporaneous. Indeed, the synchroneity of the time of construction of the two moraines is proven by the coincidence of the measured ^{10}Be concentrations at each site (Ivy-Ochs 1996). This is true irregardless of the 'actual' ^{10}Be production rate. This indicates that the measurement of cosmogenic isotope concentrations is a powerful tool for determining synchroneity of glacial events world-wide. In this way one could circumvent the problem of calibration of the radiocarbon time scale (Hajdas et al. 1995) outside of the tree ring curve (Kromer & Becker 1993).

Conclusions

The moraine complex we investigated at Julier Pass provided an ideal test. It was classified as Egesen based on its morphology and its ELA drop. Radiocarbon dates (though not specifically at this site) indicate that Egesen moraines formed during the Younger Dryas. Therefore, the age of the Egesen complex at Julier Pass was generally accepted to be Younger Dryas, but until now there has been no method of directly dating it. Our exposure age of 11,100 from six different boulders confirms that this moraine complex formed during the Younger Dryas. Additionally, we were able to unravel spatially and temporally close events such as the two events of this double-walled moraine. The outer moraine formed at the very onset of the Younger Dryas at 11,800 while the inner, more diffuse, moraine formed during the down-wasting and starving of the glacier at 10,400 years ago.

Alpine glaciers do respond to marked cold pulses in climate and do react rapidly (lag times on the order of a hundred years, Häberli 1994). Thus, they are very sensitive climate indicators and do leave a precise record of past climatic down-turns. The method of

exposure dating of moraines now allows establishment of precise chronologies for this most important terrestrial record allowing further regional and worldwide correlations (or not) of cold climatic intervals. Indeed, synchroneity of construction time of glacial features can be evaluated solely on the basis of comparison of cosmogenic isotope inventories.

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REFERENCES

ALLEY, R. B., MEESE, D. A., SHUMAN, C. A., GOW, A. J., TAYLOR, K. C., GROOTES, P. M., WHITE, J. W. C., RAM, M., WADDINGTON, E. D., MAYEWSKI, P. A. & ZIELINSKI, G. A. 1993: Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event. Nature 362, 527–529.

BORTENSCHLAGER, S. 1984: Beiträge zur Vegetationsgeschichte Tirols I. Inneres Ötztal und unteres Inntal. Ber. natw.-med. Ver. Innsbruck 71, 19 – 56.

BROECKER, W. S. 1992: Defining the boundaries of the BOA warm and YD cold isotope episodes. Quaternary Res. 38, 135–139.

BROWN, E. T., BROOK, E. J., RAISBECK, G. M., YIOU, F. & KURZ, M. D. 1992: Effective attenuation lengths of cosmic rays producing ^{10}Be and ^{26}Al in quartz: Implications for exposure age dating. Geophys. Res. Lett., 19, 369–372.

BÜHLER, C. 1983: Petrographische und geochemische Untersuchungen im Gebiet La Tscheppa-Lagrev (Julier-Bernina-Decke). Schweiz. mineral. petrogr. Mitt. 63, 457–477.

BURGA, C. A. 1988: Swiss vegetation history during the last 18000 years. New Phytologist 110, 581–602.

CERLING, T. E. & CRAIG, H. 1994: Geomorphology and in-situ cosmogenic isotopes. Annual Review of Earth and Planetary Sciences 22, 273–317.

CLARK, D. H., BIERMAN, P. R. & LARSEN, P. 1995: Improving in situ cosmogenic chronometers. Quaternary Res. 44, 366–376.

CORNELIUS, H. P. 1935/1950/1951: Geologie der Err-Julier-Gruppe. Erläuterungstext zur geologischen Karte. Kümmerly und Frey, Bern.

DANSGAARD, W., JOHNSEN, S. J., CLAUSEN, H. B., DAHL-JENSEN, D., GUNDESTRUP, N. S., HAMMER, C. U., HVIDBERG, C. S., STEFFENSEN, J. P., SVEINBJÖRNSDOTTIR, A. E., JOUZEL, J. & BOND, G. 1994: Evidence for general instability of past climate from a 250-kyr ice-core record. Nature 364, 218–220.

FABRYKA-MARTIN, J. T. 1988: Production of radionuclides in the earth and their hydrogeologic significance, with emphasis on chlorine-36 and iodine-129. Ph-D. thesis, University of Arizona, Tucson, USA.

FURRER, G., BURGA, C. A., GAMPER, M., HOLZHAUSER, H. P. & MAISCH, M. 1987: Zur Gletscher-, Vegetations- und Klimageschichte der Schweiz seit der Späteiszeit. Geographica helv. 42, 61–91.

GOSSE, J. C., EVENSON, E. B., KLEIN, J., LAWN, B. & MIDDLETON, R. 1995: Precise cosmogenic ^{10}Be measurements in western North America: Support for a global Younger Dryas cooling event. Geology 23, 877–880.

GROSS, G., KERSCHNER, H. & PATZELT, G. 1978: Methodische Untersuchungen über die Schneegrenze in alpinen Gletschergebieten. Z. Gletscherkunde und Glazialgeologie 12, 223–251.

HAJDAS, I., IVY, S. D., BEER, J., BONANI, G., IMBODEN, D., LOTTER, A., STURM, M. & SUTER, M. 1993: AMS radiocarbon dating and varve chronology of Lake Soppensee: 6000 to 12000 ^{14}C years. Climate Dynamics 9, 107–116.

HAJDAS, I., IVY-OCHS, S. D. & BONANI, G. 1995: Problems in the extension of the radiocarbon calibration curve (10–13 kyr BP). Radiocarbon 37, 75–79.

HÄBERLI, W. 1994: Accelerated glacier and permafrost changes in the Alps. In: Mountain Environments in Changing Climates (Ed. by BENISTON, M. et al.). London, United Kingdom, Routledge, 91–107.

HEUBERGER, H. 1966: Gletschergeschichtliche Untersuchungen in den Zentralalpen zwischen Sellrain und Ötztal. Wissensch. Alpenvereinshefte 20.

- 1968: Die Alpengletscher im Spät- und Postglazial. Eiszeitalter u. Gegenwart 19.

HOLZHAUSER, H. 1984: Zur Geschichte der Aletsch-Gletscher und des Fiescherglетschers. Phys Geogr. 13.

- 1987: Betrachtungen zur Gletschergeschichte des Postglazials. Geographica helv. 42, 80–87.

IVY-OCHS, S. 1996: The dating of rock surfaces using in situ produced ^{10}Be, ^{26}Al and ^{36}Cl with examples from Antarctica and the Swiss Alps. Ph-D. thesis, ETH-Zürich No. 11763.

IVY-OCHS, S., SCHLÜCHTER C., KUBIK, P. & BEER, J. 1995: Das Alter der Egesenmoräne am Julierpass. Geowissenschaften 13, 313–315.

JÄCKLI, A. 1970: Die Schweiz zur letzten Eiszeit. Atlas der Schweiz 6. Eidgenössische Landestopographie, Wabern-Bern.

JOHNSEN, S. J., CLAUSEN, H. B., DANSGAARD, W., FUHRER, K., GUNDESTRUP, N. S., HAMMER, C. U., IVERSEN, P., JOUZEL, J., STAUFFER, B. & STEFFENSEN, J. P. 1992: Irregular glacial interstadials recorded in a new Greenland ice core. Nature 359, 311–313.

KERSCHNER, H. 1978: Untersuchungen zum Daun- und Egesenstadium in Nordtirol und Graubünden (methodische Überlegungen). Geogr. Jber. aus Österreich 36, 26–49.

- 1982: Zeugen der Klimageschichte im oberen Radurschltal – Alte Gletscherstände und Blockgletscher in der Umgebung des Hohenzollernhauses. Alpenvereinsjahrbuch 1982/83 (Z.d.D.u.ö.A.V. 107, 1982), 23–27.

- 1985: Quantitative paleoclimatic inferences from lateglacial snowline, timberline and rock glacier data, Tyrolean Alps, Austria. Z. Gletscherkunde und Glazialgeologie 21, 363–369.

KINZL, H. 1929: Beiträge zur Geschichte der Gletscherschwankungen in den Ostalpen. Z. Gletscherkunde 17, 66–121.

KOHL, C. P. & NISHIIZUMI, K. 1992: Chemical isolation of quartz for measurement of in-situ produced cosmogenic nuclides. Geochim. Cosmochim. Acta 56, 3583–3587.

KROMER, B. & BECKER, B. 1993: German oak and pine 14C calibration, 7200 BC to 9400 BC. Radiocarbon 35, 125–135.

KURZ, M. D. & BROOK, E. J. 1994: Surface exposure dating with cosmogenic nuclides. In: Dating in Exposed and Surface Contexts (Ed. by BECK, C.). University of New Mexico Press, Albuquerque, 139–159.

KÜTTEL, M. 1979: Pollenanalytische Untersuchungen zur Vegetationsgeschichte und zum Gletscherrückzug in den westlichen Schweizer Alpen. Ber. Schweiz. Botan. Ges. 89, 9–62.

LAL, D. 1991: Cosmic ray labelling of erosion surfaces: in situ nuclide production rates and erosion models. Earth Planet. Sci. Lett. 104, 424–439.

LANDMANN, G., REIMER, A., LEMCKE, G. & KEMPE, S. 1996: Dating late glacial abrupt climate changes in the 14,570 years long continuous varve record of Lake Van, Turkey. PALAEO 122, 107–118.

LISTER, G. S. 1988: A 15,000-year isotopic record from Lake Zürich of deglaciation and climatic change in Switzerland. Quaternary Res. 29, 129–141.

MAISCH, M. 1981: Glazialmorphologische und gletschergeschichtliche Untersuchungen im Gebiet zwischen Landwasser- und Albulatal (Kt. Graubünden, Schweiz). Phys. Geogr. 3.

- 1982: Zur Gletscher- und Klimageschichte des alpinen Spätglazials. Geographica helv. 37, 93–104.

- 1987: Zur Gletschergeschichte des alpinen Spätglazials: Analyse und Interpretation von Schneegrenzdaten. Geographica helv. 42, 63–71.

- 1992: Die Gletscher Graubündens. Rekonstruktion und Auswertung der Gletscher und deren Veränderungen seit dem Hochstand von 1850 im Gebiet der östlichen Schweizer Alpen (Bündnerland und angrenzende Regionen). Teil A: Grundlagen-Analysen-Ergebnisse. Phys. Geogr. 33.

- 1995: Gletscherschwundphasen im Zeitraum des ausgehenden Spätglazials (Egesen-Stadium) und seit dem Hochstand von 1850 sowie Prognosen zum künftigen Eisrückgang in den Alpen. In: Gletscher im ständigen Wandel (Ed. by SCHWEIZERISCHE GLETSCHERKOMMISSION). Zürich, Switzerland, vdf Hochschulverlag AG ETH Zürich, 47–63.

MANGERUD, J., ANDERSEN, S. T., BERGLUND, B. E. & DONNER, J. J. 1974: Quaternary stratigraphy of Norden, a proposal for terminology and classification. Boreas 3, 109–127.

MASARIK, J. & REEDY, R. C. 1995: Terrestrial cosmogenic-nuclide production systematics calculated from numerical simulations: Earth Planet. Sci. Lett. 136, 381–396.

MAYR, F. & HEUBERGER, H. 1968: Type areas of Late Glacial and Post Glacial deposits in Tyrol, Eastern Alps. In: Glaciation of the Alps (Ed. by RICHMOND, G.M.), University of Colorado Studies, Series in Earth Sciences 7, 143–165.

MÜLLER, H. N. 1984: Spätglaziale Gletscherschwankungen in den westlichen Schweizer Alpen (Simplon-Süd und Val de Nendaz, Wallis) und im nordisländischen Tröllaskagi-Gebirge (Skidadalur). Küng Verlag (Näfels).

MÜLLER, H., KERSCHNER, H. & KÜTTEL, M. 1981: Gletscher- und vegetationsgeschichtliche Untersuchungen im Val de Nendaz (Wallis) – ein Beitrag zur alpinen Spätglazialchronologie. Z. Gletscherkunde und Glazialgeologie 16, 61–84.

NISHIIZUMI, K., WINTERER, E. L., KOHL, C. P., KLEIN, J., MIDDLETON, R., LAL, D. & ARNOLD, J. R. 1989: Cosmic ray production rates of ^{10}Be and ^{26}Al in quartz from glacially polished rocks. J. Geophys. Res. 94, 17,907–17,915.

PATZELT, G. 1972: Die spätglazialen Stadien und postglazialen Schwankungen von Ostalpengletschern. Ber. deutsch. Botan. Ges. 85, 47–57.

PATZELT, G. & BORTENSCHLAGER, S. 1973: Die postglazialen Gletscher- und Klimaschwankungen in der Venedigergruppe (Hohe Tauern, Ostalpen). Z. Geomorph. N.F., Suppl. Bd. 16, 25–72.

PENCK, A. & BRÜCKNER, E. 1901/1909: Die Alpen im Eiszeitalter. Tauchnitz, Leipzig.

PHILLIPS, F. M., BOWEN, D. Q. & ELMORE, D. 1994: Surface exposure dating of glacial features in Great Britain using cosmogenic chlorine-36: preliminary results. Mineral. Mag. 58A, 722–723.

PHILLIPS, F. M., ZREDA, M. G., ELMORE, D. & SHARMA, P. 1996: A reevaluation of cosmogenic ^{36}Cl production rates in terrestrial rocks. Geophys. Res. Lett. 23, 949–952.

SCHLÜCHTER, C. 1988: The deglaciation of the Swiss Alps: A paleoclimatic event with chronological problems. Bull. Assoc. franç. étude Quarternaire, 1988–2/3, 141–145.

SENARCLENS DE GRANCY, W. 1958: Zur Glazialgeologie des Ötztales und seiner Umgebung. Mitt. geol. Ges. Wien 49, 257–313.

SUTER, J. 1981: Gletschergeschichte des Oberengadins: Untersuchung von Gletscherschwankungen in der Err-Julier-Gruppe. Phys. Geogr. 2.

ZREDA, M. G. 1994: Development and calibration of the ^{36}Cl surface exposure dating method and its application to the chronology of Late Quaternary glaciations. Ph.D. thesis New Mexico Institute of Mining and Technology, Soccorro.

ZREDA, M. G. & PHILLIPS, F. M. 1994: Surface exposure dating by cosmogenic Chlorine-36 accumulation. In: Dating in Exposed and Surface Contexts (Ed. by BECK, C.). University of New Mexico Press, Albuquerque, 139–159.

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