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Earthquake-Triggered Landslides in Austria – Dobratsch Revisited

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5 Text-Figures, 1 Table

Österreichische Karte 1 : 50.000
Blatt 200

Kärnten
Dobratsch
Erdbeben

Contents

Durch Erdbeben ausgelöste Massenbewegungen in Österreich – Der Dobratsch-Bergsturz

Zusammenfassung

Ein starkes Erdbeben erschütterte im Jahr 1348 den Süden Kärntens und den Friaul. Das Ausmaß des Schadens im Gailtal und das gleichzeitige Auftreten von mehreren Bergstürzen am Dobratsch in Kärnten führten dazu, das Epizentrum in der unmittelbaren Nähe zu vermuten, nämlich in Villach. Die jüngsten Erfahrungen nutzend wird versucht, das Erschütterungspotential des größten Bergsturzes davon abzuschätzen, wobei sich zeigte, dass ein Bergsturz keinen weiteren auslösen und auch keine Schäden durch generierte Erschütterungen herbeiführen kann. Zusätzlich ist festzustellen, dass, selbst wenn der Zeitpunkt einer Hangbewegung mit einem Erdbeben zusammenfällt, dies noch nicht bedeuten muss, dass dort, wo die Hangrutschung stattfand, sich auch das Epizentrum des Erdbebens befindet.

Betrachtet man Österreich, so zeigt sich, dass die Wahrscheinlichkeit von erdbebenausgelösten Bergstürzen in jenen Bereichen am größten ist, wo auch die Erdbebengefährdung laut Österreichischer Baunorm am höchsten ist. Das südliche Kärnten zählt zu dieser Zone aufgrund der häufigen Erdbebeneinwirkungen aus dem Friaul.

Abstract

In 1348 a destructive earthquake occurred near the Austrian/Italian border. The extent of damage in Austria – and the simultaneous occurrence of several landslides at the Dobratsch massif in Southern Carinthia in Austria – gave rise to the assumption, that the epicentre must have been in Austria, probably nearby Villach. Utilising recent experience of ground shaking from landslides, and judging the potential of landslides, we find that the additional greound shaking from seismically triggered – even extremely large – landslides is insufficient to contribute to the local damage, or trigger other landslides nearby. In addition, we may conclude, that even if a landslide is clearly associated with an earthquake in time, its location does not necessarily reflect the earthquake's epicentre.

Looking at Austria, we find, that regions of the highest potential of earthquake triggered landslides coincide with the zones of highest earthquake hazard, as stipulated in the Austrian Building Code. The southern part of Carinthia also belongs to this zone, where landslides can be triggered by remote strong earthquakes which originate in nearby Friuli.

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1. Introduction

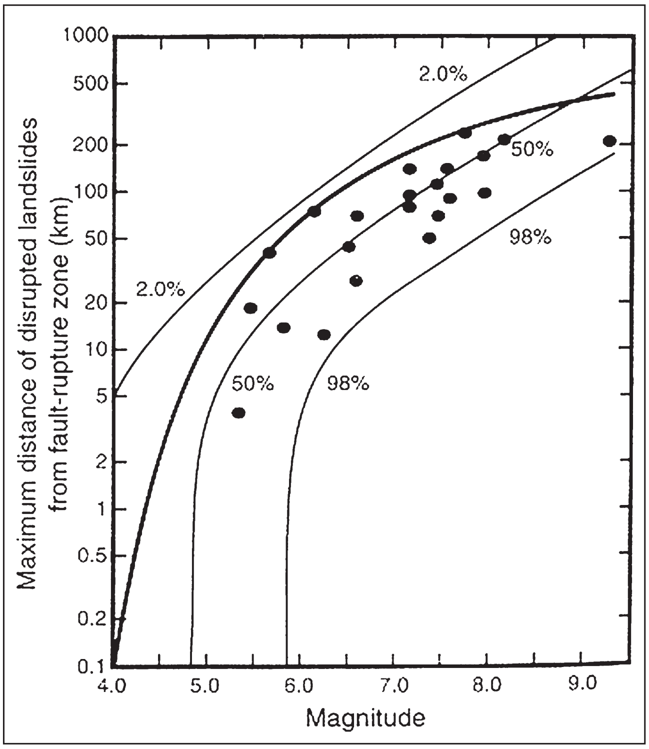
The mountainous territory of Austria is frequently subjected to larger mass movements such as landslides and rock falls. In addition, earthquakes are capable to add to this hazard by triggering additional mass movements, which would have occurred anyway but at a much later stage, when erosion further destabilized the respective rock mass. The massive landslides at the Dobratsch mountain in 1348 are one of the most impressive examples, where both factors – ground conditions and ground shaking – led to disastrous effects. The following paragraphs try to address inherent questions:

1) Where is the potential of landslides being triggered by earthquakes the highest in Austria?
2) How large can ground vibrations be caused by mass movements and which effects would they have?
3) Are landslides capable of triggering each other by means of emitted ground vibrations?

2. The Potential of Triggering Landslides by Earthquakes in Austria

Epicentral distance and the magnitude of an event govern its severity and determine how wide-spread these effects actually can be observed. Earthquakes in the Alps have the potential of reaching seismic magnitudes above 6 (e.g. Friuli 1976) and thus may cause landslides up to distances of 50 km from the epicentre (see Text-Fig. 1, from HARP & WILSON, 1995). Whether or not a slope can resist the ground motions depends mainly on the current geomechanical properties of the ground and on its water saturation (SASSA et al., 1991). To judge the potential, where earthquake-induced landslides are possible in Austria, we employ empirical estimates of ground shakings and theoretical concepts.

In 1970 Arturo ARIAS proposed a way to determine objectively the intensity of shaking by measuring the acceleration of transient seismic waves. The time-integral of the square of the ground acceleration (ARIAS, 1970)



Text-Fig. 1.
Distance of landslides versus magnitude, based on an ARIAS intensity of 0.11 m/s (HARP & WILSON, 1995).

$$I_A = \frac{\pi}{2g}\int_0^{T_d} a(t)^2 dt \tag{1}$$

with g = acceleration due to gravity and T_d = duration of signal above threshold, for practical reasons. Theoretically the integral should be infinite.

became known as "ARIAS intensity" which represents the square root of the energy per mass thus having units of "m/s".

This intensity must not be confused with the macroseismic intensity scale (GRÜNTHAL, 1998), which describes the subjective intensity of shaking as reported by people and building damage.

Table 1.
Some ARIAS intensities and their meaning (HARP & WILSON, 1995).

IA-Minimum value	Category	Description
0.11 m/s	I	Falls, disrupted slides, avalanches
0.32 m/s	II	Slumps, block slides, earth flows
0.54 m/s	III	Lateral spreads and flows

Since ARIAS intensity values have been found to be typical for certain effects in nature (Table 1), we will try to convert these values into local macroseismic intensity degrees in order to delineate regions in Austria, where this potential is existing. This can be done by substituting the magnitude in HARP & WILSON's (1995) formula derived from larger earthquakes

$$\log I_A = M_w - 2\log R - 4.1 - 0.5\,P \tag{2}$$

(M_w = moment magnitude; P = deviation from mean value in units of standard deviations)

by the magnitude derived from the scaling law of SHEBALIN (1958), which links the macroseismic epicentral intensity "I_0" to the seismic magnitude "M"

$$M = {}^2/_3\, I_0 + 2.3\log z - 2 \tag{3}$$

and substituting the term of the epicentral intensity by the macroseismic local intensity based on SPONHEUER's formula (1960)

$$I_0 = I_{local} + 3\log {}^R/_z + 1.3\,\alpha\,(R–z) \tag{4}$$

in which "R" represents the hypocentral distance and "z" is the focal depth, both in units of km. The attenuation coefficient "α" is the usually ranging from 0.001 to 0.004 with a typical value at the lower bound of 0.001/km.

In the original work, SHEBALIN derived his magnitude from surface waves. However, experience has shown, that the relation also holds for events of small magnitudes in the Eastern Alps. Therefore the stated relation has been used in Austria for more than 15 years to estimate intensities from measured local magnitudes.

In what follows, we set $M_w = M$ and $P = 0$, and we find that

$$I_{local} = 1.5\log I_A + 9.15 – 0{,}45\log z – 1.3\,\alpha\,(R–z) \tag{5}$$

Equation (5) demonstrates, that deeper earthquakes (z = 10 km) lead only to slightly smaller macroseismic intensities (~0.5°, which is also the inherent inaccuracy of macroseismic intensities) when compared with shallow ones (e.g. z = 1 km), while their ARIAS intensity is kept constant. This difference depends also very slightly on the epicentral distance which can be ignored in this context for the focal depth *"z"* is usually <10 km. In general one can state, that the ARIAS intensity serves indeed as a good indicator for the otherwise subjectively determined macroseismic intensity. Hence, an ARIAS-shaking intensity value of 0.11 m/s results in an integer value of the macroseismic intensity of 7. This intensity corresponds roughly to an effective ground acceleration in excess of 1 m/s^2.

The 10 % probability of exceedance in 50 years of such accelerations coincides with zone 4 of the Austrian Building Code for Earthquake Resistant Design "ÖNORM B 4015" (2002) based on LENHARDT (1996). Because *"P"* in equation (2) was chosen "0", equation (5) represents an average conversion from the macroseismic intensity to the ARIAS intensity – or in other words, the macroseismic intensity of 7 or the associated effective ground acceleration of 1 m/s^2 can be exceeded in 50 % of the cases.

Accordingly, areas with ground accelerations in excess of 1 m/s^2 in Text-Fig. 2 are exposed to an earthquake-related landslide hazard of approximately 5 % in 50 years, given that geological conditions permit mass movements. For macroseismic intensities smaller than degree 7, empirical relations other than equation (2) (e.g. TRAVASAROU et al., 2003) and consequently equation (5) should be used.

Regions in Austria, which can be exposed to such relatively high accelerations are – according to current knowledge – Namlos and Innsbruck in Tyrol, the Mürz-Valley in Styria and the southern part of the Vienna Basin as well as possibly Scheibbs in Lower Austria. The region of the Katschberg is also thought of being an additional area of high ground movement potential, for the 1201-earthquake has been relocated and its epicentre was shifted from Murau to the region of the Katschberg (HAMMERL, 1995). The southern part of Carinthia also qualifies as area of landslides potentially triggered by earthquakes from the northern part of Italy – Friuli.

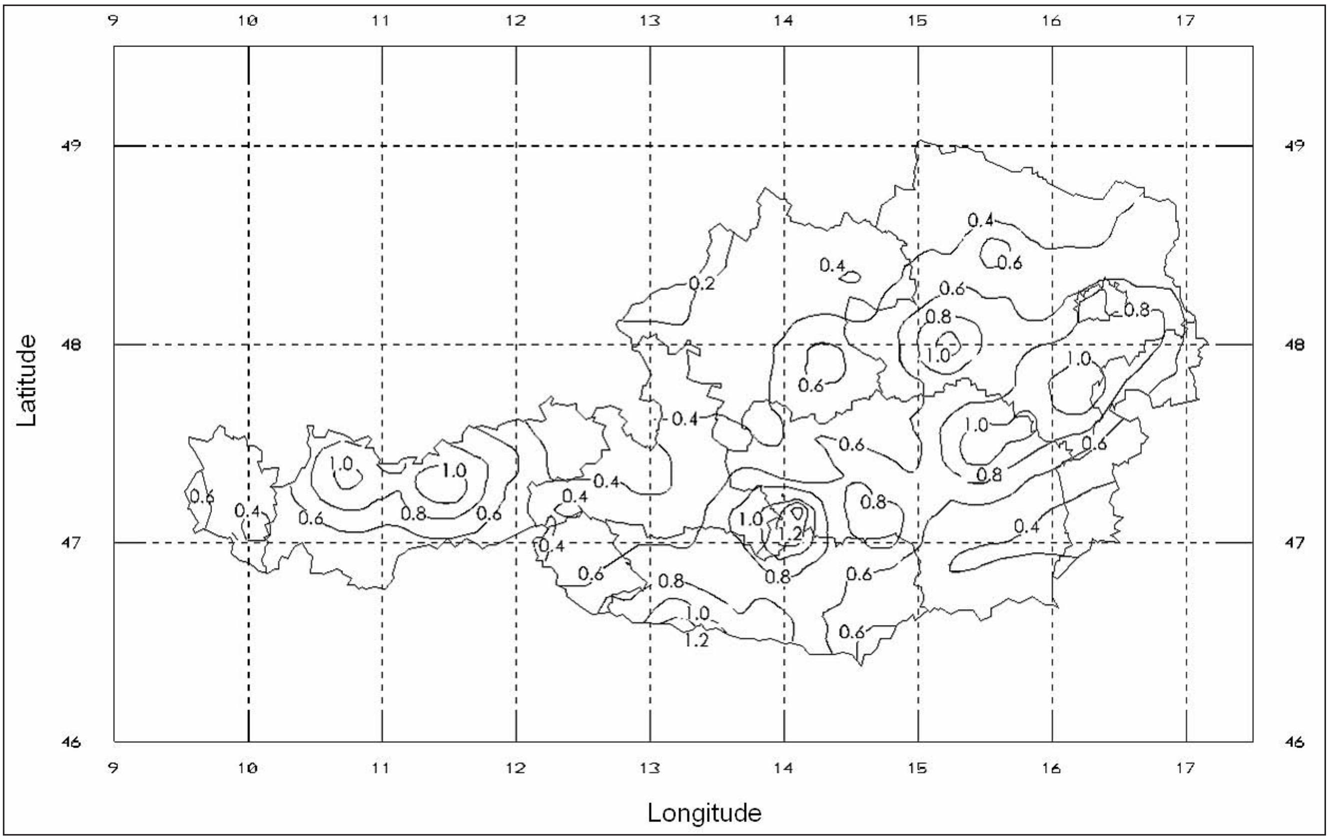
3. The Dobratsch Mountain in 1348

The earthquake in 1348, which occurred near the Austrian/Italian border, was accompanied by a number of landslides, few kilometres from the town of Villach in the province of Carinthia. During the course of this event, a number of villages were flooded by the river Gail which became dammed up by the landslide from the Dobratsch massif thus adding to the damage caused by the earthquake.

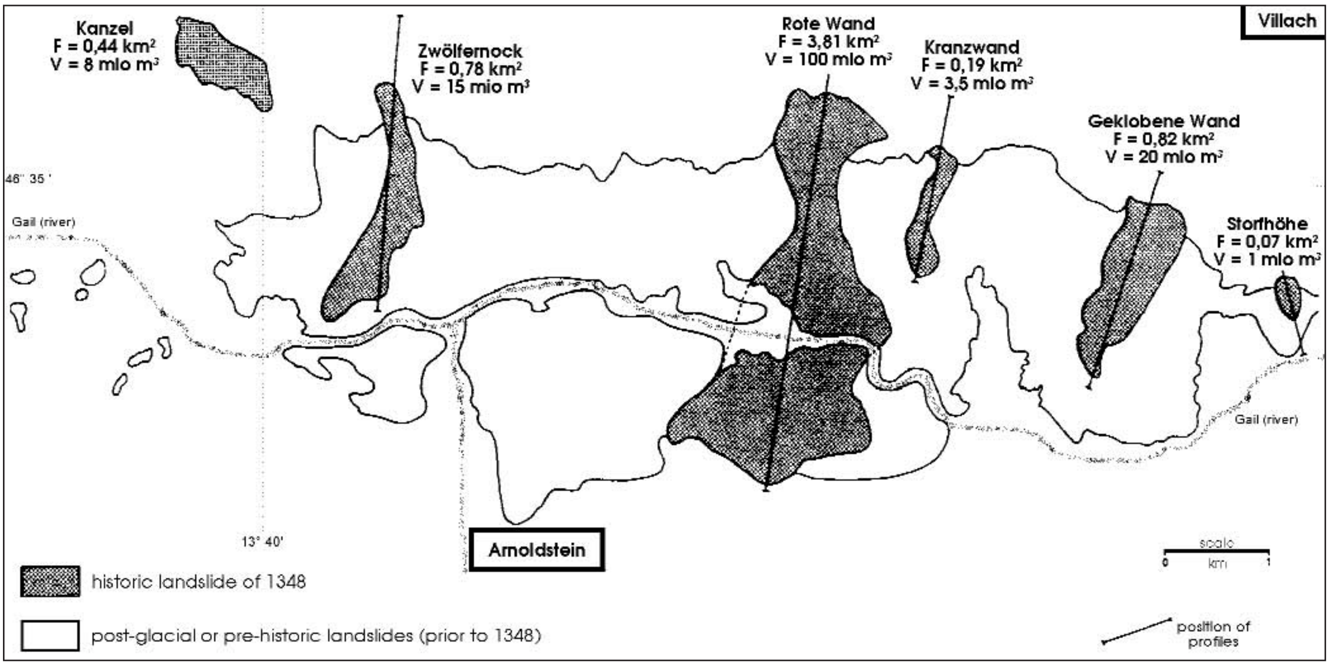
The "Dobratsch" mountain range was frequently subjected to landslides in post-glacial/pre-historical times (BRANDT, 1981). In 1348, however, actually six landslides and rock falls occurred simultaneously along the southern side of the Dobratsch, which constitutes a highly jointed geological complex. One of the landslides – along the so-called "Rote Wand" – involved by far the largest volume (Text-Fig. 3). Hence, further calculations concentrate on this particular landslide, for all other landslides and rock falls along the Dobratsch in 1348 can be regarded as less significant. The occurrence of the landslide at the Dobratsch – and the associated dramatic consequences – led to the early assumption (e.g. SIEBERG, 1940), that the epicentre must have been nearby the landslide in Carinthia, most likely near the town of Villach. This interpretation was questioned later (HAMMERL, 1992) by means of studying all available historical records of descriptions of the damage.

Landslides can involve several mechanisms, which contribute differently to ground shaking. Four phases are usually distinguished (MEISSL, 1998):

1) Toppling,
2) Bouncing, after the toppling or accelerating phases,
3) Sliding, mainly at the beginning and at the end of the blocks movement, which are mainly subjected to
4) Rolling.



Text-Fig. 2.
Effective ground accelerations in Austria with 10 % probability of exceedance in 50 years (LENHARDT, 1996).



Text-Fig. 3.
Location of six landslides associated with the earthquake of 1348 (after BRANDT, 1981).

Quite often landslides lack the primary phase, but consist of sliding and rolling phases only. Rolling does not really contribute to the emission of kinetic energy in terms of ground vibrations due to the small rolling friction. Toppling, bouncing and sliding make up for the majority of emitted seismic signals, of which "toppling" – or the impact of falling blocks in the starting phase – is the prime source. After impact, only 15 to 25 % of kinetic energy is left over (BROILLI, 1974) for the other phases of movement which gain momentum while blocks are travelling down the slope. The main part of kinetic energy is consumed by fracturing the blocks during the impact.

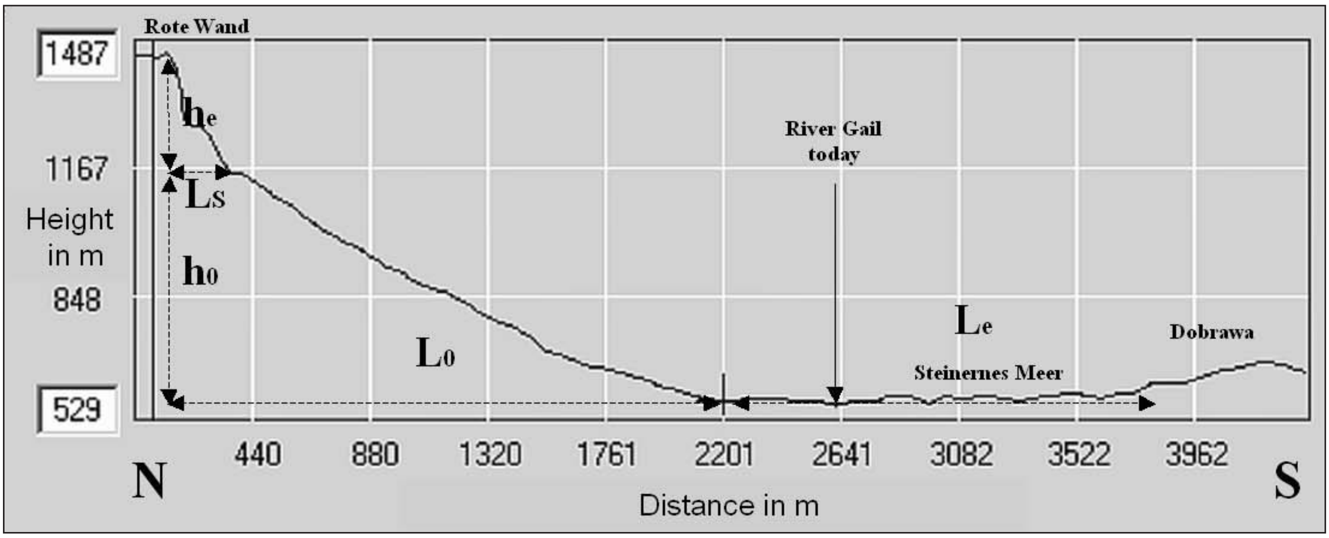
Only a fraction of this energy is transmitted into the ground making itself felt. How big this fraction is depends on conditions of the ground, onto which the blocks fall. Grass and soil damp the impact, and the local slope angle determines the partition of energy transmitted vertically into the ground, and how much energy is left for bouncing.

When looking at the section of the "Rote Wand" landslide (Text-Fig. 4), it becomes apparent, that the river bed of the "Gail" shifted in 1348. In addition, several parameters of the landslide of 1348 can be determined now:

L_o = horizontal length of slope after HEIM (1932) = 2080 m
L_e = excessive travel distance = 1590 m
L_s = length for shadow angle calculation = 320 m
L = travel distance = Lo + Le = 3670 m
h_o = height up to highest point of slope = 600 m
h_e = excessive height up to highest point of initial slide = 300 m
h = slide height = ho + he = 900 m

From these parameters, one derives several slope angles, which can be used to gain a better picture of the process involved:

α_S = shadow angle after EVANS & HUNGR (1988, 1993) $= \arctan (h_o / (L - L_s)) = 10.2°$
α_T = angle of travel distance = $\arctan (h / L) = 13.8°$
α_G = inclination of line connecting gravitational centres prior and after landslide = $\arctan (h_o + h_e / 2) / (L_o + L_e/2 - L_s/2) = 13.9°$



Text-Fig. 4.
Section of the "Rote Wand" landslide and parameters mentioned in the text. Dotted line denotes possible section prior to landslide, based on requirements for initiating the movement (x and y axis are not to scale).

α_A = average inclination of slope, down slope angle
= arctan (h_o / (L_o – L_s) = 18.8°
α_H = angle after HEIM (1932) = arctan (h / L_o) = 23.4°

Assuming a volume *"V"* of 100 million cubic metres for the sliding rock-mass and a density of 2600 kg/m^3 (VON HÜTSCHLER, 1981) results in a rock mass *"m"* of 2.6·10^{11} kg. The excessive length *"L_e"* and the horizontal slope length (L_0) can now be compared with regressions found in literature. Based on TIANCHI (1983) one finds an excessive length (in m) from the involved volume *"V"* (in m^3) using

(6) $$L_e = 10^{(0.812 + 0.322 \log (V))}$$

of 2432 m which does not compare well with the observed 1590 m. Using a volume of 30 million m^3 as mentioned by TILL (1907), we get 1658 m, however.

In addition, SCHEIDEGGER (1973) proposed a relationship between the total length *"L"* and the volume *"V"* (in m^3)

(7) $$L = h / \tan \alpha_H + 4.99 \, (V^{0.32})$$

giving a reasonable value of 3892 m which is close to the observed 3670 m. The first part of equation (7) is simply the horizontal extent of the slope *"L_0"*, while the second part of the equation expresses the excessive travel distance *"L_e"*, which is solely a function of the displaced volume and independent of the slope angle.

(8) $$\tan \alpha_A = 1.31 - 0.13 \log_8 V$$

An empirical relation between the average slope angle *"α_A"* and the volume *"V"* (in m^3) is given by (after BRANDT, 1981)

(9) $$\alpha_0 = 2 \, \alpha_A$$

whereas the initial slope angle "α_0" is related to the average slope angle *"α_A"* by
resulting in 15° for the average slope angle – which falls short by 3.8° of the actual down slope angle *"α_A"* of 18.8°, and 30° for the initial slope angle. It should be noted, that equation (9) is only valid, if the slope section can be reasonably approximated by a circle, however.

(10) $$v_{max} = \sqrt{2gL_0(\sin \alpha_A - \tan \alpha_G \cos \alpha_A)}$$

(11) $$t_{ges} = 2\,L \,/\, v_{max}$$

Letting the friction coefficient be *"$\tan \alpha_G$"*, g = 9.81 m/s^2 and using the average slope inclination *"α_A"*, we can estimate the maximum velocity with which the debris moved downhill, and the time *"t_{ges}"* it took to complete the landslide (see also UHLIR & SCHRAMM, 2003):

Hence, the maximum velocity was likely to be 60 m/s (215 km/h) at the bottom of the slope, and the whole process took approximately 120 seconds.

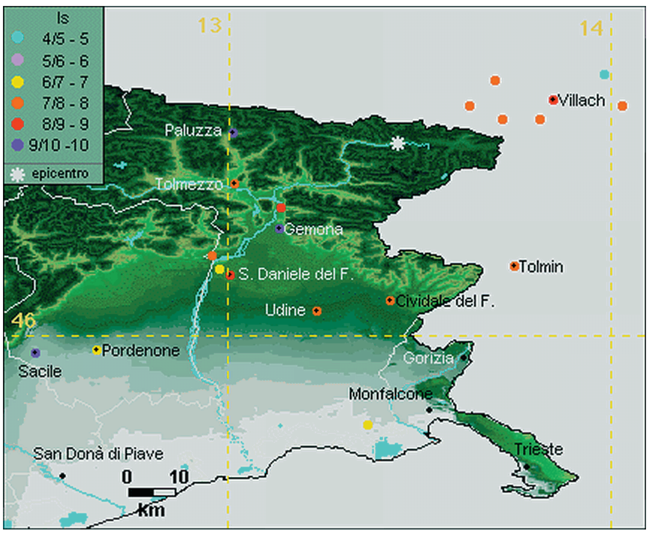
4. Earthquake Triggering

It has been shown empirically, that seismic events of M6.5 have the potential of triggering landslides up to 100 km from the epicentre (e.g. MASSANAT, 1987; HARP & WILSON, 1995), which is equivalent to a macroseismic intensity of "VI", if the focal depth does not exceed 10 km. GRÜNTHAL (1993) even states an intensity of "V" as the lower limit for possibly triggering landslides. Besides, high peak accelerations of short-duration and high-frequency have not been found to trigger large landslide masses with volumes exceeding 100 000 m^3. The triggering of massive landslides requires a low-frequency ground motion of long duration (HARP & WILSON, 1995). This might contribute to the empirical 50 %-limit of exceedance probability for disrupted landslides, which reaches as far as approx. 50 km for a magnitude M6.5-earthquake (KEEFER & WILSON, 1989).

Hence, the location of the 1348-epicentre could very well have been in Friuli, thus providing the likely background for triggering the mentioned landslide along the heavily jointed and disintegrating geological structure of the Dobratsch, which has to be regarded geomechanically highly unstable (VON HÜTSCHLER, 1981). An epicentre in Friuli is further substantiated by the lack of seismic activity along the Periadriatic Lineament, which follows the Gail valley, and the amount of reported damage which centres in Friuli, especially around Paluzza and Monte Croce Carnico. Even Carnia, Tolmezzo and Gemona del Friuli suffered extensive collapses, involving most houses and some churches (GUIDOBONI & COMASTRI, 2005; HAMMERL, 1992). MONACHESI & STUCCHI (1997) visualized the data (Text-Fig. 5) and proposed an epicentre between Pontebba and Tarvisio, possibly somewhere between Malborghetto and Camporosso.

Whether earthquakes are capable of triggering landslides depends on many factors, despite the obvious distance dependent attenuation of earthquake ground motions. VOIGHT & PARISEAU (1978) mention two earthquakes in 1969 in Hope in Canada of magnitude M3, which apparently were accompanied by landslides. Much earlier, the same region was affected by a much stronger earthquake of magnitude M7.5 in 1872. There is no evidence of landslides in the latter case, however. And again, the Friuli-earthquakes in 1976 did not trigger landslides along the Dobratsch massif as well, although ground motions appear to have reached accelerations of 1 m/s^2 locally, resulting in building damage in South Carinthia (LITSCHER & STROBL, 1977). This phenomenon seems to be characteristic for landslides, for most landslides occur without being triggered by an earthquake – and quite often earthquakes do not trigger landslides at all. A slope needs to be at the brink of failure already, if an earthquake should be able to destabilize it – or in other words – rock conditions must have dramatically deteriorated locally.

Information regarding local ground motions near a landslide are usually sparse and hinges on the deployment of seismic instrumentation (e.g. HARP & WILSON, 1995; GLAWE & MOSER, 1993). Besides, the evaluation of seismically triggered landslides turns out to be difficult as seismologists – once they are aware of the real cause – refrain



Text-Fig. 5.
Interpretation of the 1348-epicentre (MONACHESI & STUCCHI, 1997).

from stating a seismic magnitude. If they are lucky and one of their stations was close enough and recorded the event, it will give the time and the record of ground motions at the station. But they know such a "magnitude" would not adhere to the classic definition of a tectonic earthquake magnitude, which implies a certain stress drop, signal duration, frequency content and a pre-defined mechanism. Nevertheless, seismic records are very valuable when documenting landslides provided enough seismic stations covering the surroundings are available to gain a good picture of the seismic pattern (amplitude, frequency content, duration).

On the other hand, most landslides do occur without dynamic triggering from an earthquake, but due to reduced friction along the sliding surface, reduced suction – or increased pore pressure – of the involved layers. This was also recognised by BRAND (1992), who suggested that the role of seismicity as the cause for slope failure is often overemphasised world-wide, but they do occur in some cases, however. Despite inherent unknowns like such as short-term weather conditions, paleo-landslides serve as important indicators of possible paleo-seismicity (CROZIER, 1991), and attempts have even been made to map the landslide hazard, incorporating additional parameters, such as displacements, in Italy (DEL GAUDIO et al., 2003).

5. Ground Vibrations Generated by a Landslide

Judging the shaking potential of landslides, observations from Randa in Wallis/Switzerland in 1991 can be utilised in this context (SCHINDLER et al., 1993). Two massive landslides with well established volumes of 10 and 20 million m^3 caused macroseismic intensities between "II" and "III", meaning that people just noticed the rumbling at the village of Randa in the narrow valley leading to Zermatt. These dislodged volumes are comparable with those from the landslide at the Dobratsch in 1348, for TILL (1907) stated only 30,000.000 m^3 instead of 100,000.000 m^3 (VON HÜTSCHLER, 1981) for the landslide at the "Rote Wand". Such low macroseismic effects – corresponding to ground accelerations of less than 0,1 m/s^2 (1 % g) – are indicative, that only a tiny fraction of the kinetic energy available is transformed into seismic waves because the whole rock mass disintegrates into blocks of different sizes, each of which is generating relatively small seismic signals – depending on bouncing, rolling and sliding – on its own. Only the largest chunks of rock will be able to cause notable vibrations, and their respective mass is much smaller than the total stated volume „V" of the landslide, which appears to be unsuited as basic input-parameter for estimating associated ground vibrations and even subsequent damage to buildings. Hence, the landslide at the "Rote Wand" in 1348 was incapable of triggering the other landslides along the Dobratsch nor could it have contributed to the widespread damage to buildings due to vibrations associated with the process of the landslide itself – except for secondary effects such as damming up the river Gail and culminating in the flooding of numerous villages.

6. Summary

Dynamic loads brought about by seismic waves from earthquakes have a limited potential in triggering mass movements, which can be probabilistically estimated and localized. Introducing a method which allows to quantify shaking levels, regions in Austria which are thought to be prone to earthquake triggered landslides, could be delineated. The main parameter, which contributes to slope instabilities appears to be the degree of water saturation of the surface layer due to excessive rainfall, however.

The Dobratsch mountain was subjected to a number of landslides during the past. In 1348 massive and well documented landslides took place, which occurred during an earth tremor. The damage was widespread in Friuli and in the southern parts of Carinthia. Since all landslides in 1348 along the Dobratsch occurred simultaneously and they could not trigger each other, a common – and low-frequency – source of ground motion of similar vibration amplitude is necessary. Such a source is likely to be an earthquake few tens of km apart from the Dobratsch, and not in the Gail valley. An epicentre in nearby Friuli – which puts it closer to the majority of villages, from which major damage was reported – meets this requirement.

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