The Influence of Ice-Pressure on P-Wave Velocity in Alpine Low-Porosity Rocks: A Modified Time-Average Model

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Theoretically, refraction seismics are incapable of differing between frozen and unfrozen low-porosity rocks, which constitute most Alpine rock walls, since pore infill determines the changes in p-wave velocity [*McGinnis et al. 1973*]. The very opposite may be true. Here, we show the significance of the ice-pressure-induced matrix velocity increase in low-porosity rocks while velocity increase of the pore infill is negligible. Hence, the applicability of refraction seismics in low-porosity rock walls is confirmed. We present (1) the data of p-wave measurements of 22 different alpine rocks, (2) determine the increase of matrix velocity and (3) the decrease of anisotropy due to ice pressure and (4) modify Timur's [*1968*] time-average equation for alpine rocks by implementing (lithological referenced) matrix velocity changes.

Permafrost affects most high-latitude and many high-altitude regions. Permafrost distribution and dynamics have been investigated with geophysical methods since the early 1970s; the use of seismic field and laboratory measurements is a standard approach. Several laboratory studies have focused on Arctic high-porosity rocks and soils and various seismic models have been derived from the results while only Timur's [*1968*] models can be applied to rocks. In Timur's 2-phase time-average equation the increase of p-wave velocity due to freezing is based on the velocity increase of the pore infill (v_i) while matrix velocity (v_m) remains constant:

$$\frac{1}{v} = \frac{\Phi}{v_i} + \frac{1-\Phi}{v_m} \qquad (1)$$

where Φ is porosity. All tested rocks possess high porosities; however, alpine rockwalls consist of low-porosity rocks. McGinnis et al. *[1973]* implemented Timur's results in

$$\Delta v_p = \frac{\Phi - 0.0363}{0.0044} \qquad (2)$$

where Δv_p is p-wave velocity increase. Equation 2 denies applicability of seismic approaches to low-porosity rocks. Pore shape, cracks and fractures determine seismic anisotropy ("induced anisotropy") and correspond to stress, while anisotropic minerals and textural-structural characteristics cause "intrinsic anisotropy" and are stress-independent. Anisotropy is defined according to Johnston and Christensen [*1995*]

$$A = \frac{v_{\max} - v_{\min}}{v_{\max}} \qquad (3)$$

where v is the p-wave velocity parallel and perpendicular to cleavage or bedding.

Methods

We sampled 22 decimeter-large rock specimens with a natural texture of more than 100 micro-fissures and effective porosities lower than 6 % from alpine rock walls, rock glaciers and talus slopes from Switzerland, Germany, Austria, France and Svalbard. Rock samples include 14 metamorphic rocks (mostly schists and gneiss), 4 magmatic rocks (plutonites and volcanites) and 4 sedimentary rocks (e.g. sandstones and carbonates).

All samples were saturated, effective porosity and degree of saturation were determined (see Draebing & Krautblatter (subm.) for details). P-wave velocities were measured parallel and perpendicular to cleavage or bedding in a temperature range from 25°C to -15°C in a WEISS WK 180/40 high-accuracy climate chamber. 2-3 calibrated thermocouples were used to monitor continuously rock temperature; p-waves were generated with a Geotron ultrasonic transducer and measured with a Fluke Scopemeter. Matrix velocity V_m is calculated for frozen and unfrozen status according to

$$V_m = \frac{(1-\Phi)}{\left(\frac{1}{v} - \frac{\Phi}{v_i}\right)} \qquad (4)$$

and change of matrix velocity ΔV_m is assessed according to

$$\Delta V_m = \frac{(V_{mf} - V_{ms})}{V_{ms}} \qquad (5)$$

where V_{mf} is matrix velocity in the frozen status and V_{ms} is matrix velocity in the saturated status. The change of anisotropy ΔA due to freezing is calculated according to

$$\Delta A = A_s - A_f \qquad (6)$$

where A_s is the anisotropy after 48 h saturation and A_f is the anisotropy when frozen.

Results

Effective porosities of all samples are lower than 6 %; only the Tuffeau Limestone derived from non-alpine location possesses effective porosities of 45 % (Table 1). After 48 h saturation the degree of saturation of all samples is higher than 91% and pre-conditions for cryostatic pressure build-up upon volumetric expansion is achieved. Due to freezing, p-wave velocities increases by 7-78 % parallel to cleavage or bedding and 11-166 % in perpendicular direction. We plotted p-wave velocities increase due to freezing against the mean effective porosity to assume the dependence of p-wave velocity increase from porosity. For detailed results of the measurements parallel to cleavage or bedding see Draebing & Krautblatter (subm.).

Results for perpendicular measurements are shown in Fig. 1. The black line is the linear regression by McGinnis et al. [*1973*] based on measurements by Timur [*1968*] which is incaple of explaining p-wave velocity increase. Due to solution weathering and vugular pores, carbonate rocks are not plotted against porosity.

Here, p-wave velocity is dominated by matrix velocity increase (Table 1) as shown in Eq. 4 and Eq. 5. Pore ice pressure due to freezing reduces induced anisotropy from the saturated to the frozen status. Anisotropy decrease is calculated according to Eq. 6. In 15 of 22 samples anisotropy is reduced due to the developing ice pressure.

Table 1. Effective porosity, matrix velocity increase parallel and perpendicular to cleavage or bedding and anisotropy of gneiss (G), other metamorphic rocks (O), schists (S), plutonic (P) and volcanic (V) rocks, clastic (Cl) and carbonate (Ca) rocks; error values indicate mean deviation.

Rock	Porosity	parallel	perpendicular	Anisotropy
	Φ [%]	ΔV_m [%]	ΔV_m [%]	ΔA [%]
G	0.97± 0.04	5.08± 4.08	9.24± 2.23	2.92
O	1.14± 0.13	13.50± 4.46	8.95± 4.51	-3.31
S	1.48± 0.50	13.65± 7.76	61.52±45.18	15.24
P	1.43± 0.55	14.44± 1.24	14.56± 0.44	-0.35
V	3.24± 0.21	26.38± 6.33	11.81± 4.80	0.72
Cl	5.62± 0.41	23.90±20.96	17.49±13.07	-0.25
Ca	23.54±21.63	59.44± 9.93	168.53±62.00	26.96

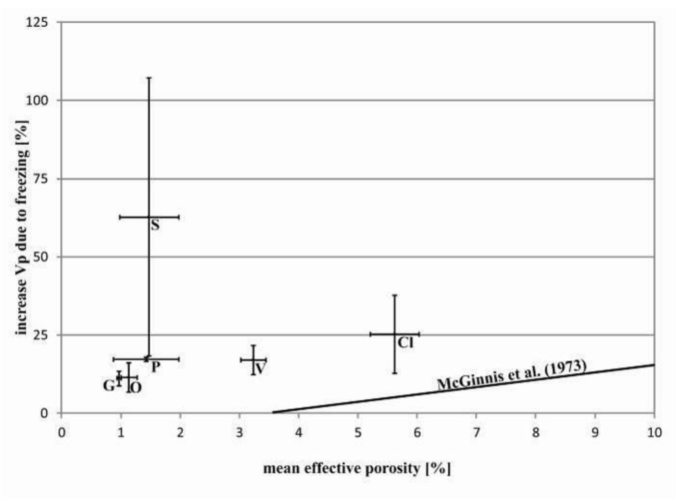

Figure 1. Percentage p-wave velocity increase due to freezing plotted against mean effective porosity for gneiss (G), other metamorphic rocks (O), schists (S), plutonic (P) and volcanic (V) rocks and clastic rocks (Cl) perpendicular to cleavage, the black line indicates the linear regression based on McGinnis et al. (1973); error bars indicate the mean deviation.

Discussion and Conclusion

P-wave velocity increase is affected by porosity but also by pore form, the degree of fissuring and ice pressure development. Pore form determines pressure susceptibility and ice effects as well as pore linkage which affects the saturation. The development of pores, fissures and fractures which provide the space and control the effects of confined ice growth is determined by the weathering history. Volumetric expansion and/or ice segregation cause ice growth which is restricted by the rigid matrix of low-porosity rocks and causes development of high levels of stress [*Matsuoka 1990*]. The combined effect of increasing matrix velocity and decreasing anisotropy result from intrinsic stress generation originated from ice pressure building. Our laboratory set up of a quasi-closed system and saturated samples are analogous to natural conditions but favor volumetric expansion more than ice segregation due to cooling rates of 6°C/h [*Matsuoka 1990*].

(1) P-wave velocity increases from 7.3 ±3.7 % for gneiss to 78.7 ±7.0 % for carbonate rocks parallel to cleavage or bedding and from 11.1 ±2.4 % for gneiss to 166.0 ±56.9 % for carbonate rocks in perpendicular direction. These increases are too high to be explained by Eq. 2. Porosity is not the relevant determinant of low-porosity bedrock.

(2) Parallel to cleavage or bedding, matrix velocity increase reaches from 5.1 ±4.1 % for gneiss to 59.4 ±10.0 % for carbonate rocks; perpendicular measurements show matrix velocity increasing from 9.0 ±4.5 % for other metamorphic rocks to 168.5 ±62.0 % for carbonate rocks.

(3) Anisotropy decreases up to 45 %, resulting from crack closure due to ice pressure in 15 of 22 rock samples.

(4) Based on our results we modify Eq. 2 with a pressure-induced variable m. We use lithology as a proxy for pore form and assume an elevated level of stress due to cryostatic pressure:

$$\frac{1}{v} = \frac{\Phi}{v_i} + \frac{1-\Phi}{v_m} * \frac{1}{m} \qquad (7)$$

where

$$m = 1 + \Delta V_m \qquad (8)$$

The proposed values of m can be derived from Table 1 for the specific lithology group.

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