P-wave velocity changes in freezing hard low-porosity rocks: a laboratory-based time-average model

D. Draebing and M. Krautblatter
Department of Geography, Bonn, Germany

Correspondence to: D. Draebing (daniel.draebing@giub.uni-bonn.de)

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Abstract. P-wave refraction seismics is a key method in permafrost research but its applicability to low-porosity rocks, which constitute alpine rock walls, has been denied in prior studies. These studies explain p-wave velocity changes in freezing rocks exclusively due to changing velocities of pore infill, i.e. water, air and ice. In existing models, no significant velocity increase is expected for low-porosity bedrock. We postulate, that mixing laws apply for high-porosity rocks, but freezing in confined space in low-porosity bedrock also alters physical rock matrix properties. In the laboratory, we measured p-wave velocities of 22 decimetre-large low-porosity (< 10 %) metamorphic, magmatic and sedimentary rock samples from permafrost sites with a natural texture (> 100 micro-fissures) from 25 °C to −15 °C in 0.3 °C increments close to the freezing point. When freezing, p-wave velocity increases by 11–166 % perpendicular to cleavage/bedding and equivalent to a matrix velocity increase from 11–200 % coincident to an anisotropy decrease in most samples. The expansion of rigid bedrock upon freezing is restricted and ice pressure will increase matrix velocity and decrease anisotropy while changing velocities of the pore infill are insignificant. Here, we present a modified Timur’s two-phase-equation implementing changes in matrix velocity dependent on lithology and demonstrate the general applicability of refraction seismics to differentiate frozen and unfrozen low-porosity bedrock.

1 Introduction

Most polar and many mountainous regions of the earth are underlain by permafrost, making them especially sensitive to climate change (IPCC, 2007; Noguès-Bravo et al., 2007). Permafrost is a thermally defined phenomenon referring to ground that remains below 0 °C for at least two consecutive years (NRC-Permafrost-Subcommitee, 1988). Permafrost is not synonymous with perennially frozen underground due to freezing point depression resulting from solutes, pressure, pore diameter and pore material (Krautblatter et al., 2010; Lock, 2005). Ice develops in pores and cavities (Hallet et al., 1991) and affects the thermal, hydraulic and mechanical properties of the underground. Climate Change can degrade permafrost and, thus, alters permafrost distribution. In mountainous regions, rockwalls with degrading permafrost are considered to be a major hazard due to rockfall activity and slow rock deformation (Gruber and Haeberli, 2007; Krautblatter et al., 2012).

Surface-based geophysical methods represent a cost-effective approach for permafrost characterization (Harris et al., 2001). The application of geophysical methods has a long tradition in permafrost studies (Akimov et al., 1973; Barnes, 1965; Ferrians and Hobson, 1973; Scott et al., 1990). Hauck and Kneisel (2008a) and Kneisel et al. (2008) provide an overview about geophysical methods suitable for permafrost monitoring in high-mountain environments. In contrast to direct temperature measurements in boreholes, geophysical methods provide only indirect information about permafrost occurrence. On the other hand, geophysical methods are non-invasive, provide spatial 2-D/3-D information and are also applicable in instable fractured rock. Frozen ground changes the properties of underground materials, the degree of change depends on water content, pore size, pore water chemistry, sub-surface temperature and material pressure (Scott et al., 1990). In field applications, the most prominent geophysical parameters for the differentiation between frozen and unfrozen underground are electrical resistivity.
and compressional wave velocity (Hauck, 2001). Alpine rock cliffs in permafrost regions mostly consist of hard low-porosity rocks (< 10%), according to Tiab and Donaldson’s (2004) definition, and the applicability of electrical and seismic methods to these is yet unclear. While the relationship between electrical resistivity and frozen low-porosity bedrock has been investigated by Krautblatter et al. (2010), this article will focus on the applicability of p-wave refraction seismics to low-porosity bedrock.

The p-wave velocity of freezing rocks was investigated in the laboratory mostly using polar high-porosity (> 10%) sedimentary rocks (Dzhurik and Leshchikov, 1973; King, 1977; Pandit and King, 1979; Pearson et al., 1986; Remy et al., 1994; Sondergeld and Rai, 2007; Timur, 1968). Only few studies included low-porosity (< 10%) sedimentary rocks (Pearson et al., 1986; Timur, 1968), igneous rocks (Takeuchi and Simmons, 1973; Toksöz et al., 1976) and metamorphic rocks (Bonner et al., 2009). Early laboratory studies demonstrated compressional and shear wave velocity increases in freezing bedrock (King, 1977; Timur, 1968). Seismic velocities increase at sub-zero temperatures until they reach a plateau when most of the pores are frozen and the unfrozen water content is negligible (Pandit and King, 1979; Pearson et al., 1986). P-wave velocity of freezing rocks is controlled by the original water-filled porosity, i.e. the velocity corresponds to the changing proportion of frozen and unfrozen pore water content (King et al., 1988). In that sense, saline pore water increases the unfrozen pore water content at a given temperature (Anderson and Morgenstern, 1973; Tice et al., 1978) and flattens the otherwise sharp p-wave velocity increase when freezing (Pandit and King, 1979). Some authors observed hysteresis effects between ascending and descending temperature runs and assumed supercooling of the pore water during the descending temperature run as a reason (King, 1977; Nakano et al., 1972).

These findings have been transferred to field applications of p-wave velocity refraction seismics to various sedimentary landforms in polar environments (Bonner et al., 2009; Harris and Cook, 1986; King, 1984; Kurfurst and Hunter, 1977; Roethlisberger, 1961; Zimmerman and King, 1986) and to rock glaciers (Barsch, 1973; Hausmann et al., 2007; Ikeda, 2006; Musil et al., 2002), to bedrock (Hauck et al., 2004) and to talus slopes (Hilbich, 2010) in mountainous regions. Akimov et al. (1973) note the discrepancy between seismic laboratory and field investigations. Due to different ambient settings, the comparison of small-scale laboratory results to large-scale field applications is complicated. These include a high rate of cooling, a non-representation of the stressed state of material as found in field conditions, supercooling and the time required for transition into ice in laboratory studies.

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Wyllie et al. (1956) developed a time-average equation

\[ \frac{1}{v} = \frac{\Phi}{v_1} + \frac{1 - \Phi}{v_m}, \]

where \( v \) is the measured velocity, \( v_1 \) is the velocity of the liquid inside the pore space, \( v_m \) is the matrix velocity and \( \Phi \) is the porosity, based on measurements of sandstone (0.02 < \( \Phi < 0.32 \)) and limestone samples (0.001 < \( \Phi < 0.18 \)). The time-average equation requires a relative uniform mineralogy, fluid saturation and high effective pressure (Mavko et al., 2009). To fulfil the seismic ray assumption of the time-average equation the wavelength should be small compared with typical pore and grain size, respectively, and the pores and grains should be arranged as homogenous layers perpendicular to seismic ray path (Mavko et al., 2009). Due to larger size and more heterogeneous distribution of vugular, i.e. secondary solution-related, pores in carbonate rocks, p-wave velocities of carbonate rocks show less dependency on porosity and the time-average equation underestimates the p-wave velocities (Wyllie et al., 1958). The two-phase model of Timur (1968) modified the Eq. (1) to frozen states,

\[ \frac{1}{v} = \frac{\Phi}{v_1} + \frac{1 - \Phi}{v_m}, \]

where \( v_1 \) is the velocity of ice in the pore space. Timur (1968) extended Eq. (2) to a three-phase time-average equation:

\[ \frac{1}{v} = \frac{(1 - S)\Phi}{v_1} + \frac{S_i\Phi}{v_i} + \frac{1 - \Phi}{v_m} \]

with \( S_i \) is the relative fraction of pore space occupied by ice. Equation (2) and Eq. (3) were tested for sandstone (0.13 < \( \Phi < 0.42 \)), carbonate (0.15 < \( \Phi < 0.47 \)) and shale samples (0.04 < \( \Phi < 0.10 \)). McGinnis et al. (1973) deduced that the relative p-wave velocity increases upon freezing \( \Delta v_p \) [%] versus porosity is

\[ \Delta v_p = \frac{\Phi - 0.0363}{0.0044} \]

based on a linear regression of Timur’s (1968) measurements; a formula that implies that there are no p-wave velocity changes below 3.6% porosity. This relation was only used as an interpretation tool for their field measurements and possesses no validity for low-porosity rocks. Hauck et al. (2011) extended Timur’s (1968) equation to 4 phases and weighted the p-wave velocities of the components by their volumetric fractions:

\[ \frac{1}{v} = \frac{f_1}{v_1} + \frac{f_m}{v_m} + \frac{f_i}{v_i} + \frac{f_a}{v_a} \]

\[ f_1 + f_m + f_i + f_a = 1 \text{ and } 0 \leq f_i, \ f_m, \ f_i, \ f_a \leq 1 \]

where \( v_a \) is the velocity of air, \( f_i \) is the volumetric fraction of liquid water, \( f_i \) is the volumetric fraction of ice, and \( f_m \) is the volumetric fraction of rock.
Pores react to an increasing confining pressure according to their shape: spheroidal pores deform and become thinner while spherical pores decrease in volume (Takeuchi and Simmons, 1973; Toksöz et al., 1976). P-wave velocity will increase due to decreasing porosity if the confining pressure does not surpass the damage threshold and porosity increase due to microcracking (Eslami et al., 2010; Heap et al., 2010; Wassermann et al., 2009). In measurements with high confining pressures, the effect of pores is negligible but the effects of cracks become more important (Takeuchi and Simmons, 1973). In frozen rocks, the ice pressure effect is most pronounced for spheroidal “flat” pores or cracks (Toksöz et al., 1976).

Pore shape, cracks and fractures also determine seismic anisotropy next to anisotropic mineral components and textural-structural characteristics such as bedding and cleavage (Barton, 2007; Lo et al., 1986; Thomsen, 1986; Vernik and Nur, 1992; Wang, 2001). The two latter causes are referred to as intrinsic anisotropy and cannot decrease as a result of pressure (Barton, 2007; Lo et al., 1986; Thomsen, 1986). In contrast, “induced anisotropy” through pores, cracks and fractures corresponds to stress. Stress increase due to loading can preferentially close pre-existing microcracks perpendicular to stress direction and decreases anisotropy (Eslami et al., 2010; Heap et al., 2010; Wassermann et al., 2009). However, stress increase can also lead to preferential opening of axially orientated microcracks (Eslami et al., 2010) or microcrack generation due to threshold surpassing (Heap et al., 2010; Wassermann et al., 2009), which then enhances anisotropy. The anisotropy $A$ is defined as

$$A = \frac{v_{\text{max}} - v_{\text{min}}}{v_{\text{max}}},$$

where $v_{\text{max}}$ is the faster velocity of both compressional waves parallel and perpendicular to cleavage or bedding and $v_{\text{min}}$ is the slower velocity (Johnston and Christensen, 1995).

We postulate that p-wave velocity measurements in low-porosity rocks could become an important method for the monitoring of Alpine rock wall permafrost. This study aims at (1) measuring the p-wave velocity increases in low-porosity rocks, (2) evaluating the increase of matrix velocity due to ice pressure, (3) describing the alteration of seismic anisotropy due to changes of induced pore pressure and (4) incorporating this matrix velocity increase in the time-average equation.

## 2 Methodology

We tested 20 Alpine and 2 Arctic rock specimens between 1.8 and 25 kg sampled from several permafrost sites (see Table 1 for details). We used large rock specimens with a statistical distribution of $>100$ fissures, cracks and cleavages in a sample to cope with natural bedrock heterogeneity (Akomov et al., 1973; Jaeger, 2009; Matsuoka and Murton, 2008). All samples were immersed in water under atmospheric conditions until full saturation indicated by a constant weight was achieved ($W_s$). The free saturation method resembles the field situation more closely than saturation under vacuum conditions (Krus, 1995; Sass, 2005) but probably includes air bubbles and can complicate the interpretation. After that, the samples are dried at 105°C to a constant weight ($W_d$). The ratio of weight difference between saturated and dry weight is equal to moisture content in percentage by weight. This multiplied by the rock density is effective porosity $\Phi_{\text{eff}}$ and includes only hydraulically-linked pores (Sass, 2005). Rock density is derived from Wohlenberg (2012).

To distinguish quantitatively connected and unconnected pores will help the interpretation but necessary methods were not available. In an earlier study by Krautblatter (2009), six plan-parallel cylindrical plugs were prepared with diameter and length of 30 mm from six of the 22 samples used in this study and porosity values were measured using a gas compression/expansion method in a Micromeritics Multivolume Pycnometer 1305. These absolute porosity values are used to estimate the quality of the effective porosity values.

All 22 samples were immersed again for 48 h under atmospheric conditions and the saturated weight $W_{48\text{h}}$ was determined. To determine the moisture conditions we calculated the degree of saturation $S_t$

$$S_t = \frac{(W_{48\text{h}} - W_d)}{(W_s - W_d)}. \quad \text{(10)}$$

Subsequently, samples were loosely coated with plastic film to protect them against drying and were cooled in a range of 25°C to $-15°C$ in a WEISS WK 180/40 high-accuracy climate chamber (Fig. 1). The cooling rate was first 7°C h$^{-1}$ until sudden p-wave velocity increase due to
Table 1. Rock samples, source, lithology, geologic description (fabric), porosity (Φ) and degree of saturation (Sr).

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location (Country)</th>
<th>Landform</th>
<th>Sort of Lithology</th>
<th>Geological description (fabric)</th>
<th>Porosity (Φ) [%]</th>
<th>Degree of Saturation (Sr)</th>
<th>Publ.</th>
</tr>
</thead>
<tbody>
<tr>
<td>A5</td>
<td>Steintaelli/Mattervalley (CH)</td>
<td>Rock wall</td>
<td>Q</td>
<td>pyritic paragneiss</td>
<td>metamorphic lamination</td>
<td>0.04 ± 0.14</td>
<td>0.99</td>
</tr>
<tr>
<td>H1</td>
<td>Matterhorn/Mattervalley (CH)</td>
<td>Rock wall</td>
<td>S</td>
<td>gneiss</td>
<td>metamorphic lamination</td>
<td>0.93 ± 0.12</td>
<td>1.00</td>
</tr>
<tr>
<td>X5</td>
<td>Zastler/Black Forest (D)</td>
<td>Scree slope</td>
<td>S</td>
<td>gneiss</td>
<td>metamorphic lamination</td>
<td>0.95 ± 0.12</td>
<td>1.00</td>
</tr>
<tr>
<td>X2</td>
<td>Pastoritz/Grosglockner (A)</td>
<td>Glacier forefield</td>
<td>S</td>
<td>serpentine</td>
<td>mixed fabric directions</td>
<td>0.97 ± 0.04</td>
<td>1.00</td>
</tr>
<tr>
<td>H2</td>
<td>Matterhorn/Mattervalley (CH)</td>
<td>Rock wall</td>
<td>S</td>
<td>amphibolite</td>
<td>mixed fabric directions</td>
<td>1.27 ± 0.16</td>
<td>0.96</td>
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<td>S1</td>
<td>Steintaelli/Mattervalley (CH)</td>
<td>Rock wall</td>
<td>Q</td>
<td>schist</td>
<td>schistite</td>
<td>1.14 ± 0.13</td>
<td>1.11</td>
</tr>
<tr>
<td>S4</td>
<td>Steintaelli/Mattervalley (CH)</td>
<td>Rock wall</td>
<td>Q</td>
<td>schist</td>
<td>planar schistosity</td>
<td>2.40 ± 0.12</td>
<td>0.94</td>
</tr>
<tr>
<td>S3</td>
<td>Steintaelli/Mattervalley (CH)</td>
<td>Rock wall</td>
<td>Q</td>
<td>quartz</td>
<td>planar quartzite</td>
<td>1.94 ± 0.10</td>
<td>0.94</td>
</tr>
<tr>
<td>X8</td>
<td>Munte/Upper Engadin (CH)</td>
<td>Rock glacier</td>
<td>S</td>
<td>gneiss</td>
<td>planar schistosity</td>
<td>1.49 ± 0.08</td>
<td>0.98</td>
</tr>
<tr>
<td>A22</td>
<td>Bliggspitz/Kaunervalley (A)</td>
<td>Rock wall</td>
<td>Q</td>
<td>mica schist</td>
<td>planar schistosity</td>
<td>1.86 ± 0.13</td>
<td>0.97</td>
</tr>
<tr>
<td>X13</td>
<td>Kitzsteinhorn (A)</td>
<td>Rock wall</td>
<td>S</td>
<td>mica schist</td>
<td>planar schistosity</td>
<td>1.56 ± 0.11</td>
<td>1.00</td>
</tr>
<tr>
<td>D2</td>
<td>Murag/Upper Engadin (CH)</td>
<td>Rock glacier</td>
<td>S</td>
<td>mica schist</td>
<td>planar schistosity</td>
<td>0.83 ± 0.06</td>
<td>0.91</td>
</tr>
<tr>
<td>C1</td>
<td>Corvatsch/Upper Engadin (CH)</td>
<td>Rock wall</td>
<td>Q</td>
<td>migmatic schist</td>
<td>planar schistosity</td>
<td>1.04 ± 0.07</td>
<td>1.00</td>
</tr>
<tr>
<td>C2</td>
<td>Corvatsch/Upper Engadin (CH)</td>
<td>Rock wall</td>
<td>Q</td>
<td>migmatic schist</td>
<td>planar schistosity</td>
<td>1.60 ± 0.15</td>
<td>0.99</td>
</tr>
<tr>
<td>M1</td>
<td>Aiguille du Midi (F)</td>
<td>Rock wall</td>
<td>Q</td>
<td>granite</td>
<td>no pronounced fabric</td>
<td>1.48 ± 0.30</td>
<td>1.00</td>
</tr>
<tr>
<td>X9</td>
<td>Gemstock (CH)</td>
<td>Rock wall</td>
<td>Q</td>
<td>granodiorite</td>
<td>plagioclastic rocks</td>
<td>2.25 ± 0.05</td>
<td>0.99</td>
</tr>
<tr>
<td>X6</td>
<td>Prag/Black Forest (D)</td>
<td>Scree slope</td>
<td>S</td>
<td>andesite</td>
<td>no pronounced fabric</td>
<td>3.03 ± 0.35</td>
<td>1.00</td>
</tr>
<tr>
<td>X7</td>
<td>Prag/Black Forest (D)</td>
<td>Scree slope</td>
<td>S</td>
<td>andesite</td>
<td>no pronounced fabric</td>
<td>3.45 ± 0.40</td>
<td>1.00</td>
</tr>
<tr>
<td>L1</td>
<td>Longeuardalen/Svalbard (N)</td>
<td>Rock wall</td>
<td>Q</td>
<td>Endalen Sandstone</td>
<td>no pronounced fabric between bedding planes (sample size)</td>
<td>1.09 ± 0.06</td>
<td>1.00</td>
</tr>
<tr>
<td>L2</td>
<td>Longeuardalen/Svalbard (N)</td>
<td>Talus slope</td>
<td>S</td>
<td>Endalen Sandstone</td>
<td>no pronounced fabric between bedding planes (sample size)</td>
<td>1.40 ± 0.14</td>
<td>1.00</td>
</tr>
<tr>
<td>A8</td>
<td>Zugspitze/Wettersteinengebirge (Da)</td>
<td>Rock wall</td>
<td>Q</td>
<td>Wetterstein Dolomite</td>
<td>no pronounced fabric between bedding planes (sample size)</td>
<td>45.16 ± 0.54</td>
<td>0.95</td>
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<tr>
<td>K1</td>
<td>Saumur/Loire Valley (F)</td>
<td>Sedimentary basin</td>
<td>Q</td>
<td>Tuffeau Limestone</td>
<td>carbonate rocks</td>
<td>25.34 ± 0.54</td>
<td>0.95</td>
</tr>
</tbody>
</table>


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Fig. 1. Laboratory measurement set up of a p-wave velocity measurement of a schist/gyrazite slate sample (S1) in parallel direction to cleavage. Drilled into the rock sample are three thermometers to monitor rock temperature.

Freezing and was then decreased to 6 °C h⁻¹ (Matsuoka, 1990). Ventilation was applied to avoid thermal layering. Two to three calibrated 0.03 °C-accuracy thermometers were drilled into the rock samples to depths between 3 and 10 cm and a spacing of approximately up to 10 cm depending on sample size. Rock temperature at different depths and spacings were measured to account for temperature homogeneity in the sample (Krautblatter et al., 2010). The p-wave generator Geotron USG 40 and the receiver were placed on flattened or cut opposite sides of the cuboid samples. The wavelength of the generator was 20 kHz to fulfill requirements of the time-average equation; dispersion of p-wave velocities due to wavelengths are negligible (Winkler, 1983). The travel time of the p-wave was picked using a Fluke ScopeMeter 192B with an accuracy of 1–2 × 10⁻⁶ s. The internal deviation induced by the measurement procedure was assessed by conducting five subsequent travel time measurements. To account for the anisotropy of the rock samples, we measured p-wave velocities in the same sample in the direction of cleavage/bedding and perpendicular to the cleavage/bedding direction. The matrix velocity $v_m$ is calculated by solving Eq. (2). The velocity of the material in the pore space $v_i$ is 1570 m s⁻¹ for water in the unfrozen status and 3310 m s⁻¹ for ice (Timur, 1968), we replaced porosity with effective porosity in the calculation. Matrix velocity is calculated for frozen (−15 °C) and unfrozen status (mean value of $v > 0$ °C) both for parallel and perpendicular to cleavage/bedding measurements according to

$$ v_m = \frac{1 - \Phi}{1 - r^2} \frac{v_i}{v} $$

(11)
The change of matrix velocity $\Delta v_m$ due to freezing is calculated according to

$$\Delta v_m = \frac{v_{mf} - v_{ms}}{v_{ms}},$$

where $v_{mf}$ is the matrix velocity in the frozen status and $v_{ms}$ is the matrix velocity in the saturated status. The change of anisotropy $\Delta A$ due to freezing will be calculated according to

$$\Delta A = A_s - A_f,$$

where $A_s$ is the anisotropy after 48 h saturation and $A_f$ is the anisotropy for frozen status.

### 3 Results

Tables 1 and 2 give an overview about measured rock samples and their rock properties and seismic velocities. Figure 2 represents the evolution of p-wave velocities dependent on rock temperature of six selected rock samples from six different lithologies.

#### 3.1 Porosities and degree of saturation

The absolute (vacuum) porosity values comprehending connected and non-connected porosity measured for 6 samples (A5, X2, S1, S3, X9, A8) by Krautblatter (2009) are compared with the effective (atmospheric pressure) porosity values comprehending only connected porosity. The absolute porosity $2.60 \pm 0.21$ % is on average 30 % higher than the effective porosity $1.72 \pm 0.12$ %, only in slate samples both were equivalent.

Rock samples are classified according to their lithology into three metamorphic, two igneous and two sedimentary rock clusters. Absolute deviations of porosity within the clusters are less than 1 % except for carbonate rock samples. After 48 h saturation, gneiss, plutonic rocks, volcanic rocks and clastic rocks show mean $S_f$ of 1.00; other metamorphic rocks (mean $S_f = 0.98$); schists (mean $S_f = 0.97$) and carbonate rocks (mean $S_f = 0.98$) are not fully saturated, but all could possibly develop cryostatic pressure on the volumetric expansion of ice in more than 91 % saturated pores (Walder and Hallet, 1986).

#### 3.2 P-wave velocities of frozen rock

P-wave velocity increases significantly as a result of freezing in all 22 samples. Supercooling causes hysteresis effects resulting in sudden latent heat release and rock temperature increase observed in 16 of 22 samples and indicated as p-wave velocity hysteresis of three rock samples (A5, X8, L2) in Fig. 2. Parallel to cleavage/bedding, p-wave velocity increase is highest in sedimentary (carbonate and clastic) rocks, followed by magmatic (volcanic and plutonic) rocks and lowest in metamorphic rocks (schists, other metamorphic rocks and gneiss) (Fig. 3a). The order remains the same perpendicular to cleavage/bedding except for schists (Fig. 3b).

#### 3.3 Porosity dependent change in p-wave velocities

Existing time-average models assume a dependence of p-wave velocity increase on porosity. We plotted the increase of p-wave velocity due to freezing measured and calculated with Eq. (2) against the mean effective porosity (Fig. 4a and b). We excluded the carbonate rocks due to their vugular pores and the constrained applicability of the time-average equation (Wyllie et al., 1958). All measured p-wave velocity increases are much higher than calculated according to Eq. (2), expected as a result of phase transition from water (1570 m s$^{-1}$) to ice (3310 m s$^{-1}$) only. Parallel to cleavage or bedding, the offset between measured and calculated results is increasing from gneiss (296 $\pm$ 205 m s$^{-1}$), schists (642 $\pm$ 314 m s$^{-1}$), other metamorphic rocks (685 $\pm$ 200 m s$^{-1}$), plutonic rocks (686 $\pm$ 0 m s$^{-1}$), clastic rocks (815 $\pm$ 683 m s$^{-1}$), to volcanic rocks (1158 $\pm$ 278 m s$^{-1}$). Perpendicular to cleavage or bedding, the offset increases from other metamorphic rocks (414 $\pm$ 210 m s$^{-1}$), gneiss (467 $\pm$ 108 m s$^{-1}$), volcanic rocks (529 $\pm$ 183 m s$^{-1}$), plutonic rocks (561 $\pm$ 41 m s$^{-1}$), clastic rocks (626 $\pm$ 474 m s$^{-1}$) to schists (1368 $\pm$ 695 m s$^{-1}$).

#### 3.4 Matrix velocity

The increase in p-wave velocity is too high to be solely explained by changes of the p-wave velocity in the pore infill as is suggested by Timur (1968). Here, the additional change in p-wave velocity is explained by the increase in matrix velocity as shown in Eq. (11) and Eq. (12). All measured rock samples show significant matrix velocity increases $v_m$ (see Table 2) due to freezing except one gneiss sample (X5). Timur
Table 2. Rock samples classified into lithological groups and seismic properties. The table shows p-wave velocity of a saturated unfrozen ($v_{ps}$) and a frozen ($v_{pf}$) sample, p-wave velocity increase due to freezing ($\Delta v_{p}$), matrix velocity of a saturated unfrozen ($v_{mf}$) and a frozen ($v_{mf}$) sample, matrix velocity increase due to freezing ($\Delta v_{mf}$), anisotropy of a saturated ($A_s$) and a frozen ($A_f$) sample and the decrease of anisotropy due to freezing ($\Delta A$).

<table>
<thead>
<tr>
<th>Sample/ Rock</th>
<th>P-wave velocity</th>
<th>Matrix Velocity</th>
<th>Anisotropy</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>parallel</td>
<td>perpendicular</td>
<td>parallel</td>
</tr>
<tr>
<td></td>
<td>$v_{ps}$ [m s$^{-1}$]</td>
<td>$v_{pf}$ [m s$^{-1}$]</td>
<td>$\Delta v_{p}$ [m s$^{-1}$]</td>
</tr>
<tr>
<td>A5</td>
<td>6281</td>
<td>6608</td>
<td>428</td>
</tr>
<tr>
<td>H1</td>
<td>5401</td>
<td>6099</td>
<td>698</td>
</tr>
<tr>
<td>X5</td>
<td>5699</td>
<td>5826</td>
<td>127</td>
</tr>
<tr>
<td>X2</td>
<td>5275</td>
<td>5873</td>
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(1968) expected no matrix velocity increase due to freezing. Figure 4 shows that Timur’s Eq. (2) underestimates the measured p-wave velocity significantly.

3.5 Anisotropy

Anisotropy $A$ is calculated according to Eq. (9) for conditions after 48 h saturation ($A_t$) and frozen conditions at $-15^\circ$C ($A_f$). Induced anisotropy due to pores, cracks and fractures can be reduced through pressure (Barton, 2007; Wang, 2001). Anisotropy alteration $\Delta A$ is calculated according to Eq. (13). In our experimental setup, pore ice pressure reduces induced anisotropy due to the closure of pores, cracks and fractures, while the confining (atmospheric) pressure remains constant. The pore pressure changes due to the phase transition from water to ice in saturated pores. Ice develops pressure through volumetric expansion and ice segregation (Matsuoka, 1990; Matsuoka and Murton, 2008). 15 of 22 samples show an anisotropy reduction due to freezing (1–45%), which is especially pronounced in slates, schists and carbonates. Seven samples show negligible ($n = 3, < 1.50\%$) or small ($n = 4, \leq 3.50\%$) increases in anisotropy when freezing. Three samples (L1, L2, M1) show a low anisotropy increase, the anisotropy of four other samples (H1, H2, X2, X7) increases slightly.

4 Discussion

4.1 Model setup and representativeness

Previous studies (McGinnis et al., 1973; Timur, 1968) explained p-wave velocity increases exclusively as an effect of porosity and infl in. We postulate, that these models apply well for soft high-porosity rocks but cannot be transferred to hard low-porosity rocks. This is due to the fact that the effects of freezing are determined by multiple factors including (i) porosity but also (ii) the pore form and the degree of fissuring and (iii) ice pressure development. Here, we try to derive a straightforward model that explains the effects of freezing in low-porosity rocks on p-wave velocity.

(i) We have tested 7 clusters or 22 specimens of low-porosity rocks. These indicate p-wave velocity increases from $518 \pm 227$ m s$^{-1}$ (other metamorphic rocks), $542 \pm 105$ m s$^{-1}$ (gneiss), $658 \pm 63$ m s$^{-1}$ (plutonic rocks), $767 \pm 164$ m s$^{-1}$ (volcanic rocks), $927 \pm 414$ m s$^{-1}$ (clastic rocks), $1427 \pm 681$ m s$^{-1}$ (schists) to $3193 \pm 927$ m s$^{-1}$ (carbonates) perpendicu lar to cleavage or bedding. Timur’s (1968) model would, respectively, anticipate p-wave velocity changes from $518 \pm 19$ m s$^{-1}$ (gneiss), $96 \pm 238$ m s$^{-1}$ (volcanic rocks), $1427 \pm 165$ m s$^{-1}$ (clastic rocks), $3193 \pm 104$ m s$^{-1}$ (carbonates) to $3193 \pm 982$ m s$^{-1}$.
Fig. 3. P-wave velocity increase of samples in percent for rock groups classified based on lithology; (A) parallel to cleavage/bedding and (B) perpendicular to cleavage/bedding; error bars indicate mean deviation.

(clastic rocks) to $58 \pm 34 \text{ m s}^{-1}$ (schists) and underestimates strongly p-wave velocity increases in low-porosity bedrock (Fig. 4b). Due to vugular pores, the time-average equation and Eq. (2) are not applicable to carbonate rocks (Wyllie et al., 1958) and we excluded them from further calculations. The offset between measured velocities and calculated velocities shows that porosity is not the dominant determinant of p-wave velocity changes in low-porosity bedrock. Calculations of p-wave velocities parallel to cleavage or bedding reflect this offset trend but are violating the seismic ray assumptions of the time-average equation and should be used with caution for parallel velocities.

(ii) Pore form is among the most important factors for seismic properties (Nur and Simmons, 1969; Toksöz et al., 1976; Wang, 2001) and the most difficult one to quantify (Wang, 2001). Pore form determines pressure susceptibility (Takeuchi and Simmons, 1973; Toksöz et al., 1976) and ice effects (Toksöz et al., 1976) while pore linkage affects the saturation. Water-saturated porosity controls p-wave velocity (King, 1977; King et al., 1988) and frost weathering (Matsuoka, 1990; Matsuoka and Murton, 2008; Sass, 2005). We assume no influence of salinity due to low solubility of rock minerals in the used specimens (Krautblatter, 2009). Hydraulically linked porosity is best described by effective porosity (Sass, 2005) and we replace porosity in Eq. (2) with effective porosity. In future studies, the pore form could be assessed by porosimetric analyses and, thus, the differentiation of connected and non-connected porosity would facilitate a quantitative interpretation. However, calculating matrix velocity with absolute porosity values would change matrix velocity only by $2 \pm 2 \%$, which is well below the accuracy within the clusters. The weathering history determines the enlargement of pores, fissures and fractures in permafrost and non-permafrost samples, and we assume that the long periglacial weathering history of high-alpine and arctic samples affects pore shape and connectivity. Previous mentioned studies mostly used high-porosity arctic specimens from Mesozoic sedimentary rocks and frost susceptibility in these low-strength rocks operates at a millimeter- to centimeter-scale (Matsuoka
and Murton, 2008). We choose decimeter-large rock samples from several Alpine and one Arctic permafrost sites instead of standard bore cores. These are derived from the surface or quarried out of rock walls, are affected by permafrost in their history, include hundreds of micro-fissures, and represent the natural texture of permafrost-affected bedrock. This reflects that properties like pore distribution, texture, fissures and fractures provide the space and determine the effects of confined ice growth in hard rock samples (Matsuoka and Murton, 2008). In hard rocks, volumetric expansion and ice segregation is restricted by the rigid matrix and ice growth in pores and fissures causes high levels of stress inside the samples.

(iii) The variation of confining pressure related to rock overburden is a long-lasting process on a millennium scale, whereas pore pressure changes steadily (Matsuoka and Murton, 2008). Frequent daily freeze−thaw cycles reach a depth of approximately 30 cm (Matsuoka and Murton, 2008) while annual cycles often reach up to 5 m and more (Matsuoka et al., 1998). In our experiment the change in matrix velocity in combination with reduced anisotropy points towards “induced anisotropy” (Wang, 2001) in pores that reflects intrinsic stress generation. The pore pressure in the connected pores presumably increases due to ice stress applied on the matrix and probably closes non-connected porosity embedded in the matrix which results in decreasing anisotropy. A surpassing damage threshold or opening of microcracks could explain anisotropy increase. The pore pressure can be generated by the ice pressure building (Matsuoka, 1990; Vlahou and Worster, 2010) due to volumetric expansion of in situ water (Hall et al., 2002; Matsuoka and Murton, 2008) and ice segregation (Hallet, 2006; Murton et al., 2006; Walder and Hallet, 1985). In the laboratory, any open system allows water migration and enables ice segregation while closed systems with water-saturated samples favour volumetric expansion (Matsuoka, 1990). Our experimental setup is a quasi-closed system; water is only in situ available due to saturation and ice can leave through pores and joints. Due to 48 h saturation, the degree of saturation reaches at least 0.91 in all samples and the threshold for frost cracking as a result of volumetric expansion is fulfilled (Walder and Hallet, 1986). According to Sass (2005) and Matsuoka (1990) our quasi-closed system and fully saturated samples could be a good analogue to natural conditions.

Cooling rates of 6 °C h−1 have been used by Matsuoka (1990) before and produce high expansion and freezing strain. Sass (2005) assumes high saturation of alpine rocks below the upper 10 cm. This is due to the fact that ice pressure is relaxed through ice deformation and ice expansion into free spaces (Tharp, 1987), ice extrusion (Davidson and Nye, 1985) and the contraction of samples was observed in the long-term due to ice creep (Matsuoka, 1990). In our system, samples cool from all outer faces which presumably act to seal the sample with ice. On the other hand, ice segregation along temperature gradients in fissured natural bedrock will cause suction up to several MPa (Murton et al., 2006; Walder and Hallet, 1985) and ice growth, and presumably cause a persistent elevated level of cryostatic stress similar to our laboratory setup.

4.2 A time-average model for low-porosity rock

Figure 4a and b show an offset which is not explainable by Eq. (2). This offset is induced by ice pressure. The way ice pressure is effective depends on the pore form of connected and non-connected pores. A quantitative analysis needs to distinguish between connected and non-connected pores. We use lithology as a proxy for pore form in our model and we assume an elevated level of stress in cryostatic systems. The pressure-induced variable m depends on lithology and is introduced as an extension of Eq. (2):

$$\frac{1}{v} = \frac{\Phi}{v_i} + \frac{1 - \Phi}{v_m} \times \frac{1}{m}$$  \hspace{1cm} (14)

where

$$m = 1 + \Delta v_m;$$  \hspace{1cm} (15)

$\Delta v_m$ is the increase of matrix velocity empirically derived from our measurements. These general conclusions referenced by rock type are preliminary and should be applied with caution since we used a restricted number of samples. For our rock samples, we propose values of $m$ of 1.09 ± 0.02 for gneiss, 1.09 ± 0.05 for other metamorphic rocks, 1.62 ± 0.45 for schists, 1.15 ± 0.00 for plutonic rocks, 1.12 ± 0.05 for volcanic rocks and 1.17 ± 0.13 for clastic rocks or, alternatively a general $m$ of 1.34 ± 0.31 (Table 2). The use of Eq. (14) enhances to differentiate between frozen and unfrozen status of low-porosity rocks and can facilitate interpretation of field data.

5 Conclusions

Here, we propose to incorporate the physical concept of freezing in confined space into empirical mixing rules of p-wave velocities and present data (1) of p-wave measurements of 22 different alpine rocks, (2) evaluate the influence of ice pressure on seismic velocities, (3) determine anisotropic decrease due to ice pressure and (4) extend Timur’s (1968) 2-phase model for alpine rocks:

(1) All tested rock samples show a p-wave velocity increase dependent on lithology due to freezing. P-wave velocity increases from 418 ± 194 ms−1 for gneiss to 2290 ± 370 ms−1 for carbonate rocks parallel to
cleavage/bedding; perpendicular measurements show an acceleration ranging from 518 ± 105 m s⁻¹ for other metamorphic rocks to 3193 ± 996 m s⁻¹ for carbonate rocks.

(2) P-wave velocity increases due to freezing are dominated by an increase of the velocity of the rock matrix while changes in pore-infill velocities are insignificant. Matrix velocity increases perpendicular to cleavage/bedding from 420 ± 221 m s⁻¹ for other metamorphic rocks to 1387 ± 717 m s⁻¹ for schists; parallel measurements reflect the matrix velocity increases perpendicular to cleavage but should be treated with caution.

(3) Anisotropy decreases by up to 45% as a result of crack closure due to ice pressure in 15 of 22 rock samples. This effect is observed especially in all samples containing planar slaty cleavage or planar schistosity.

(4) We developed a novel time-average equation based on Timur's (1968) 2-phase equation with a lithology dependent variable to increase the matrix velocity responding to developing ice pressure while freezing. This study provides the physical basis for the applicability of refraction seismics in low-porosity permafrost rocks. Due to their rigidity low-porosity bedrock cannot expand freely in response to ice pressure and, thus, matrix velocity increases. P-wave velocity increases predominantly as a result of ice pressure and to a lesser extent as a result of the higher velocity of ice than water in pores. The extension of the time average equation provides a more realistic calculation of the rock velocity and facilitates the interpretation of field data and possible permafrost distribution in alpine rock walls.

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