Chapter 1

INTRODUCTION

Dehydration processes have been proposed to be a major controlling parameter in many subduction related processes [Newton, 1990; Peacock, 1990; Davies, 1999; Abers, 2000; Stern, 2002; Schurr et al., 2003]. Particularly the weakening of the subduction channel within the mantle wedge [e.g. Gerya & Stöckhert, 2002], arc magmatism [e.g. Gill, 1981; Davies & Stevenson, 1992; Pearce & Peate, 1995] and the occurrence of intermediate-depth earthquakes [Kirby et al., 1996; Hacker et al., 2003a, b] are assumed to be caused by the release of fluids from the subducted slab. Furthermore, pore pressure excess may also be involved in tectonic processes in and below accretionary prisms [e.g. Byrne & Fisher, 1990; Moore & Vrolijk, 1992]. One of the difficulties in verifying these hypotheses is that the release and migration of fluids in the Earth is a transient process on various temporal as well as spatial scales, which is hard to observe directly with geophysical methods especially in depths exceeding 10 – 20 km. However, in the laboratory processes such as the dehydration of rocks and its relation to petrophysical properties can be studied in detail with the aid of physical *in-situ* measurements under defined pressure and temperature conditions.

The release of fluids will enhance the pore-fluid pressure in a rock till its tensile strength is exceeded and the rock fails, leading to fluid-induced seismicity. Both, the inflation of pores as well as hydrofracturing and a subsequent collapse of the fractures can be interpreted as a dynamic modification of the rock microstructure in terms of a reconstitution of pore space and variations in mineral contacts. Sound waves are highly sensitive to these microstructural variations. Thus, ultrasonic experiments are an ideal tool to study fluid–rock interactions *in-situ*.

Exogene as well as *endogene* parameters are known to affect the elastic properties of rocks. The *exogene* parameters are the pressure and temperature conditions, to which a rock is exposed. The *endogene* parameters can be subdivided into *intrinsic* factors, such as the chemical and mineralogical composition as well as the matrix structure (e.g. lattice preferred orientation of minerals) and *extrinsic* factors, which include the structure and geometry of void space (such as cracks and pores) and the sample history (e.g. stress memory). In response to the fast technical development during World War II (fast-writing oscillo-

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scopes, ultrasonic devices) numerous laboratory experiments have been performed since the 1950's, to investigate elastic wave velocities of rocks and their pressure *(P)* and temperature *(T)* dependency [e.g. Hughes & Maurette, 1956, 1957; Birch, 1960, 1961; Simmons, 1964a, b; Christensen, 1965, 1966, 1979, Kern, 1979; Kern et al. 1997; Ito & Tatsumi, 1995]. At first, sound velocities were determined as a function of the chemical and mineralogical composition of the rocks and were later supplemented by studies addressing the influence of microstructural parameters. In particular, besides the effect of structural anisotropy [e.g. Kern, 1979; Siegesmund et al., 1989; Prasad et al., 1994] the impact of the dimension and orientation of pores and cracks and their filling [e.g. O'Connell & Budiansky, 1974; Siegesmund & Vollbrecht, 1991; Hudson, 2000; Rasolofosaon et al., 2000] were studied experimentally and theoretically. Experimentally, the effect of cracks and pore pressure has mostly been related to geotechnological problems, and has therefore been examined primarily in the low pressure range $\leq$ 300 *MPa* [Nur & Simmons, 1969; Prasad & Manghnani, 1997]. In contrast, experiments conducted under high confining pressures focussed on the characterisation of *intrinsic*, matrix related, effects, which are so complex and manifold that the influence of a modified void structure at higher confining pressures is usually ignored. Thus, there are no investigations on the effect of the pore pressure on the elastic properties at high confining pressure yet, firstly, because of the technical requirements of the set-up, and secondly, because pores at greater depth in the Earth are neglible at the geological time scale [Warner, 2004]. Nonetheless, fluids are regarded to be intimately linked to a number of tectonic processes in a variety of geological settings, and hence, the study of the influence of the pore pressure on the rock properties also at high confining pressures is of particular interest to understand these transient processes.

Several mechanisms are proposed to influence the development of the pore pressure [Bredehoeft & Hanshaw, 1968; Domenico & Palciauskas, 1988; Ko et al., 1997]:

- A change in the pore volume due to mechanical or chemical compaction processes (compression of the rock, mineral dissolution – precipitation).

- The change of fluid volume, for instance by thermal expansion or

- the release of structural water bound in hydrous minerals.

The latter is regarded as the major factor controlling the pore pressure excess when water is released during dehydration. So far, especially in experiments at high pressure and temperature the samples were usually studied at drained conditions, where the fluid phases discharged during dehydration reactions of hydrous minerals were released to the atmosphere.

As fully drained systems are unlikely to occur at greater depth in nature, this study follows a different approach. According to the importance of the role of fluids in the triggering of earthquakes, which are assumed to be related to a local and temporary increase of pore-fluid pressure, the consideration of an

undrained ("closed") system under high confining pressure is of great importance to better understand the variation in physical properties of a dehydrating rock in nature. The study of the elastic properties in terms of the modification of the void structure due to the presence of supercritical fluids is thus a key for the understanding of pore pressure induced processes in the interior of the Earth.

In the present study, experiments on encapsulated rock samples (amphibolite and serpentinite) were performed. Beyond the determination of *intrinsic* elastic properties, this allows to study the influence of variations in pore fluid pressure on the microstructure (*extrinsic* properties) and thus of changes in the elastic properties of a rock. It is already known that dehydration reactions, which occur in a narrow *PT* window of several tens of °C, like those observed in gypsum, serpentinite, and zeolite and chlorite bearing rocks, are linked to a sharp decrease in P wave velocities even under drained conditions [Kern, 1982; Popp, 1994; Lebedev & Kern, 1999]. Other minerals, such as mica, and amphibole dehydrate more continuously over a broader range of *PT* conditions, which might result in a variation of the seismic signature of this process. Furthermore, deformation experiments on gypsum, chlorite as well as serpentinite at drained and undrained conditions generally reveal a dehydration-induced weakening and embrittlement of the samples, whereby the strength of the drained samples is less affected [Raleigh & Paterson, 1965; Heard & Rubey, 1966; Murrell & Ismail, 1976]. All these observations lead to the following questions:

1. Are the velocity-temperature relations of drained systems similar to those of undrained systems or do they exhibit significant differences?

2. How much pore fluid is required to have a measurable effect on the seismic velocity of an undrained rock at elevated confining pressure?

3. What effect has the thermal expansion of a fluid stored in an undrained sample?

4. Is there a difference in the seismic signature of discontinuous *(serpentinite)* and continuous *(amphibolite)* dehydration reactions?

5. Is there a relation between hydromechanical processes and the texture of a rock?

To answer these questions elastic wave velocities were determined for amphibolite and serpentinite samples, encapsulated in an impermeable jacket to achieve undrained conditions. To obtain reliable results a special experimental set-up was designed. The ultrasonical experiments were complemented by microstructural analyses of the starting materials and of the partially dehydrated samples to investigate the evolution of porosity. The results provide an *in-situ* insight into different mechanisms controlling the modification of the pore space due to the presence of supercritical pore fluids.

1.1   Elastic Properties of Homogeneous Solids

The propagation of an elastic wave through an elastic medium can be described by the infinitesimally small deformations in the transmitted body, responding to an external stress. Then, Hook's approximation states, that the strain ε_{kl} is proportional to the applied stress σ_{ij} and depends on the material properties, described for example by the constants of elasticity, C_{ijkl} [e.g. Feynman, 1963].

$$\sigma_{ij} = C_{ijkl} \cdot \varepsilon_{kl}. \tag{1.1}$$

The tensor C_{ijkl} consists of 81 components. Because of the symmetry of both the stress tensor and the strain tensor and due to the reversibility of elastic deformation the number reduces to maximum 21 independent elements. They can be further reduced depending on the symmetry of the medium [e.g. Feynman, 1963]. In the simplest case of an isotropic elastic solid two elastic moduli are sufficient to completely describe the elastic behaviour of the material, e.g. the *Lamé constants G* and λ, whereby G is the *shear modulus* or *rigidity modulus* and a measure of the resistance of the medium to shear. Eq. 1.1 then reduces to

$$\sigma_{ij} = \lambda \cdot \Delta \cdot \delta_{ij} + 2 \cdot G \cdot \varepsilon_{ij}, \tag{1.2}$$

with the *Kronecker delta* δ_{ij} equal to 1 for $i = j$ and 0 for $i \neq j$, respectively, and for small deformation the *cubical dilatation* becomes $\Delta = \varepsilon_{xx} + \varepsilon_{yy} + \varepsilon_{zz}$.

Other frequently used elastic moduli and the *Poisson's ratio* can be calculated from the two *Lamé constants*: The *incompressibility* or *bulk modulus K*, a measure for the resistance of the medium to volumetric changes, is given by

$$K = \lambda + \frac{2}{3} G. \tag{1.3}$$

Young's modulus E is defined as the resistance to elongation

$$E = \frac{3\lambda + 2G}{(\lambda + G)} G, \tag{1.4}$$

and the *Poisson's ratio* ν, defined as the ratio of the fractional lateral contraction to the fractional longitudinal extension for a small cylinder under tension at both ends, is

$$\nu = \frac{\lambda}{2(\lambda + G)}. \tag{1.5}$$

$$\frac{\Delta V}{V} = -\frac{P}{K}$$

V: volume
ΔV: volume change
P: pressure
K: bulk modulus

$$\varphi = -\frac{\tau}{G}$$

φ: shear angle
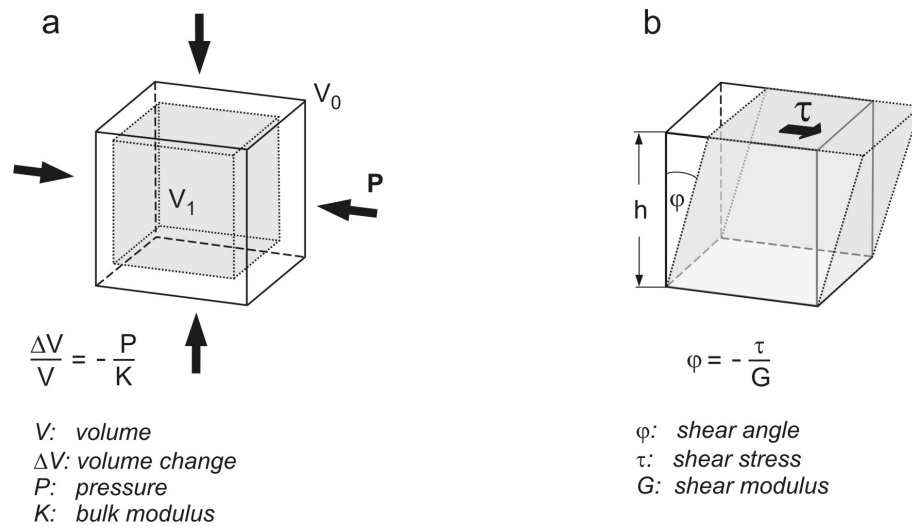τ: shear stress
G: shear modulus

Figure 1.1: *Illustration of bulk modulus (a) and shear moduls (b).*

Solving the equations of motion through a homogeneous, isotropic elastic medium and using Eq. 1.2, the velocities for the two types of planar waves can be derived in isotropic solids. The velocity for *longitudinal* waves (P waves) is

$$v_p = \sqrt{\frac{\lambda + 2G}{\rho}} = \sqrt{\frac{K + \frac{4}{3}G}{\rho}}. \tag{1.6}$$

P waves are also called compressional waves, as the particle density varies in the direction of wave propagation.

In contrast, in *transversal* waves (S waves) the particle oscillation is perpendicular to the propagation direction and involves shearing motions without any volume changes

$$v_s = \sqrt{\frac{G}{\rho}}. \tag{1.7}$$

Finally, the dimensionless *Poisson's ratio* can be expressed by V_p and V_s

$$v = \frac{1}{2}\frac{(v_p/v_s)^2 - 2}{(v_p/v_s)^2 - 1}. \tag{1.8}$$

1.2   Elastic Properties of Rocks

The elastic constants depend on pressure as well as temperature. These dependencies lead to a variation of elastic wave velocities in the case of changing pressure and temperature (*PT*) conditions, whereby *P* and *T* have an opposite effect. The *PT* behaviour has been investigated in several studies over the last decades and is elucidated examplarily on two datasets [Christensen, 1979; Kern et al., 1997] in Fig. 1.2: The increase of confining pressure provokes the closing of microcracks and a progressive reduction of pore space, which is reflected in a strong non-linear increase of the velocity in the lower pressure range (Fig. 1.2 a) [e.g. Birch, 1961; Christensen, 1965, 1966]. This behaviour is strongly related to the closure of microcracks. The flattened, linear slope above 200 MPa, on the other hand, is dominated by the intrinsic properties of the compacted rock matrix, which are in a good agreement with properties derived from appropriate averaging single crystal data [e.g., Mainprice & Humbert, 1994]. In other words, this part of the velocity–pressure relation is attributed to the pressure dependence of the elastic constants of the rock-forming minerals. In most cases, the intrinsic velocities in the lower pressure range cannot be measured directly in the laboratory and are extrapolated by linear regression of the intrinsic behaviour at elevated pressures.
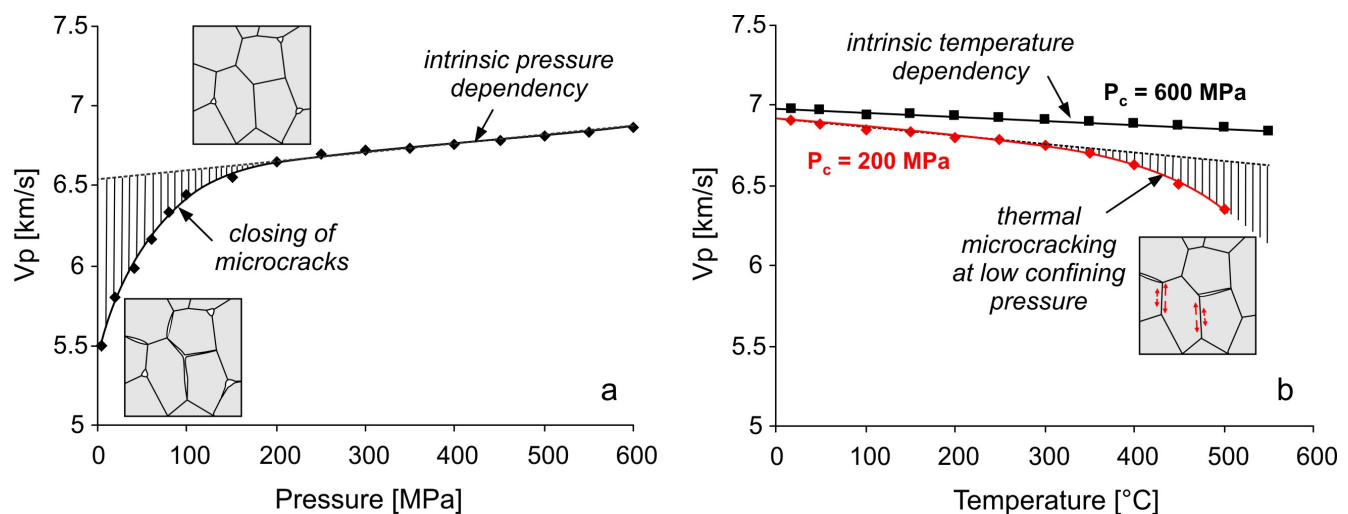

Figure 1.2: *Elastic wave velocities as a function of pressure (a) and temperature (b). Data were measured by Kern et al., 1997 (black) and Christensen, 1979 (red).*

The effect of temperature on the elastic wave velocity is illustrated in Fig. 1.2 b. The increase of temperature usually lowers the bulk and shear moduli of a material, leading to a linear decrease of the velocity–temperature relation. The development of non-linearity above 300 °C at 200 MPa in Fig. 1.2 b is attributed to thermal cracking, due to different thermal expansion coefficients and anisotropy of the rock-forming minerals. The onset of thermally induced microcracking is a function of the confining

pressure. On the basis of a series of measurements Kern [1978] concluded, that a pressure increase of about 1 MPa per 1 K temperature increase is required to suppress thermally induced cracking.

Elastic wave velocities are very sensitive to changes in the void structure of the transmitted medium caused by external parameters such as pressure, temperature, and pore fillings. The injection of a pore-fluid establishes a fluid pressure in the void system of the rock. Therefore, the effective pressure P_{eff}, defined by the confining pressure P_c and the pore-fluid pressure P_{fl}, decreases [Brandt, 1955; Biot, 1956a, b; Todd & Simmons, 1972].

$$P_{eff} = P_c - n\, P_{fl}. \tag{1.9}$$

The factor n ($0 < n \leq 1$) has been described as the effective stress coefficient or the coefficient of pore space deformation. For the simplification of $n = 1$ it can be written :

$$P_{diff} = P_c - P_{fl}, \tag{1.10}$$

where P_{diff} is called the *differential* pressure. Moreover, any change of P_{eff} or P_{diff} affects the state of microfracturing and may therefore result in a change of elastic properties of the rock. The release of structurally bonded water and trapping inside the *undrained* samples used in this study, should decrease in particular the P wave velocities significantly. To prevent the superposition of thermally and fluid induced microcracking, this study was carried out at high pressures between 750 to 900 MPa. Following the estimation of Kern [1978] thermally induced cracking should then be negligible up to temperatures of 750 to 900 °C.