University of Anbar

College of Science

Department of Applied Geology

Advanced Structural Geology

Title of the lecture

The Deformation Apparatus

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The Deformation Apparatus

A deformation experiment on a rock or a mineral can be carried out readily by placing a small sample in a vise, but when you try this experiment, you have to be careful to avoid being bombarded with randomly flying chips as the material fails. In rock deformation experiments we attempt to control the experiment a little better for the sake of the experimentalist, as well as to improve the analysis and interpretation of the results. A typical deformation apparatus is schematically shown in Figure 1. In this rig, a cylindrical rock specimen is placed in a pressure chamber, which is surrounded by a pressurized fluid that provides the **confining pressure**, P_c, on the specimen through an impermeable jacket. This experimental setup is known as a **triaxial testing apparatus**, named for the triaxial state of the applied stress, in which all three principal stresses are unequal to zero. For practical reasons, two of the principal stresses are equal. In addition to the fluid that provides the confining pressure, a second fluid may be present in the specimen to provide **pore-fluid pressure**, P_f. The difference between confining and pore pressure, $P_c - P_f$, is called the effective pressure (P_e). Adjusting the piston at the end of the test cylinder results in either a maximum or minimum stress along the cylinder axis, depending on the magnitudes of fluid pressure and axial stress. The remaining two principal stresses are equal to the effective pressure. By varying any or all of the axial stress, the confining pressure, or the pore-fluid pressure, we obtain a range of stress conditions to carry out our deformation experiments. In addition, we can heat the sample during the experiment to examine the effect of temperature. This configuration allows a limited range of finite strains, so a torsion rig with rotating plates is increasingly used for experiments at high shear strains.

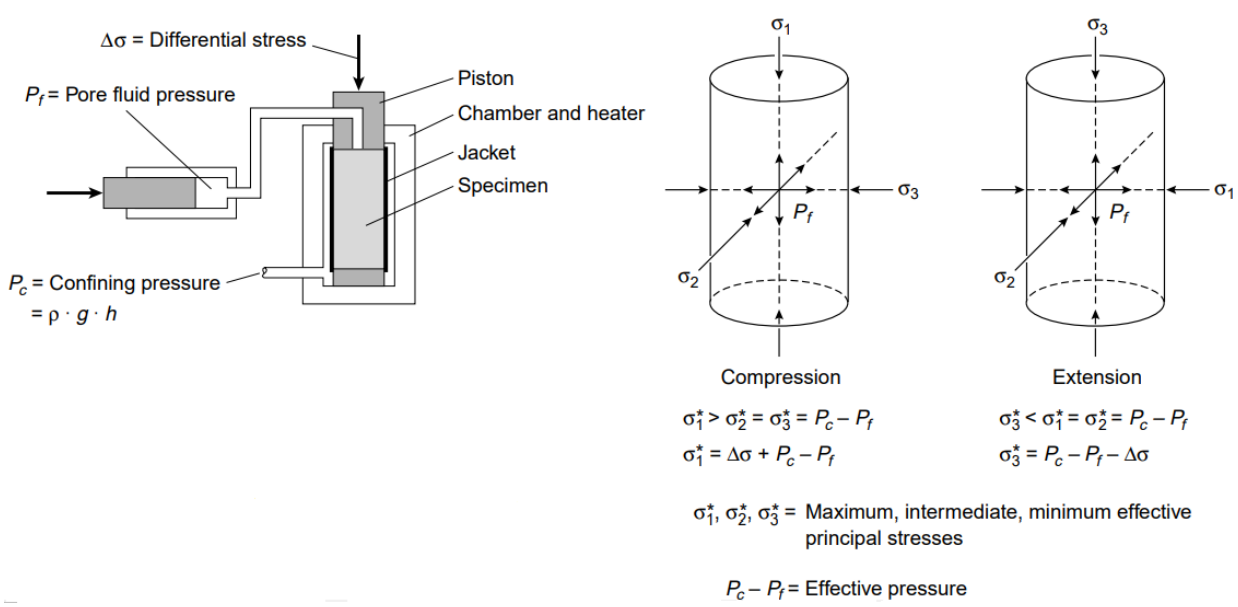

Figure 1schematic diagram of a triaxial compression apparatus and states of stress in cylindrical specimens in compression and extension tests. the values of pc, pf, and σ can be varied during the experiments.

A triaxial apparatus enables us to vary stress, strain, and strain rate in rock specimens under carefully controlled parameters of confining pressure, temperature, pore-fluid pressure, and time (that is, duration of the experiment). What happens when we vary these parameters and what does this tell us about the behavior of natural rocks? Before we dive into these experiments it is useful to briefly review how these environmental properties relate to the Earth. Both confining pressure and temperature increase with depth in the Earth (Figure 2). The confining pressure is obtained from the simple relationship

$$Pc = \rho \cdot g \cdot h$$

where ρ is the density, g is gravity, and h is depth. This is the pressure from the weight of the overlying rock column, which we call the lithostatic pressure. The temperature structure of the Earth is slightly more complex than the constant gradient of 10 K/km. At first, temperature increases at an approximately constant rate (10°C/km–30°C/km), but then the thermal gradient is considerably less (Figure 2). Additional complexity is introduced by the heat generated from compression at high pressures, which is reflected in the adiabatic gradient (dashed line in Figure 2). But if we limit our considerations to the crust and uppermost mantle, a linear geothermal gradient in the range of 10°C/km–30°C/km is a reasonable approximation. The most geologic processes occur at strain rates on the order of 10^{-14}/s, with the exception of meteoric impacts, seismic events, and explosive volcanism. In contrast to geologic strain rates, experimental work is typically limited by the patience and the life expectancy of the experimentalist. Some of the slowest experiments are carried out at strain rates of 10^{-8}/s (i.e., 30% shortening in a year), which is still four to seven orders of magnitude greater than geologic rates. Having said all this, now let's look at the effects of varying environmental conditions, such as confining pressure, temperature, strain rate, and pore-fluid pressure, during deformation experiments.

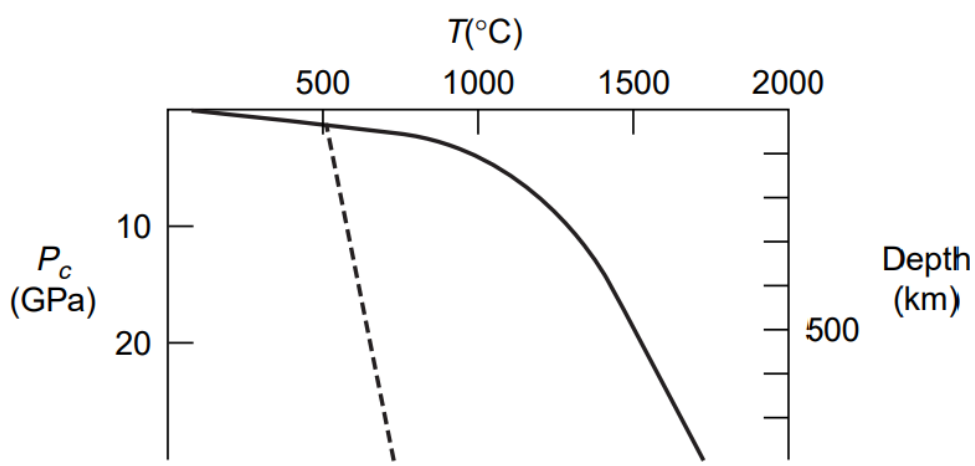

Figure 2 Change of temperature (T) and pressure (Pc) with depth. The dashed line is the adiabatic gradient, which is the increase of temperature with depth resulting from increasing pressure and the compressibility of silicates.

Confining Pressure

The confining pressure acts equally in all directions, so it imposes an isotropic stress on the specimen. When we change the confining pressure during our experiments, we observe a very important characteristic: increasing confining pressure results in greater strain accumulation before failure (Figure 1). In other words, increasing confining pressure increases the viscous component and therefore the rock's ability to flow. What is the explanation for this? If you have, the Mohr circle for stress and failure criteria give an explanation, but we will assume that this material is still to come. So, we'll take another approach. Moving your arm as part of a workout exercise is quite easy, but executing the same motion under water is a lot harder. Water "pushes" back more than air does, and in doing so it resists the motion of your arms. Similarly, higher confining pressures resist the opening of rock fractures, so any shape change that occurs is therefore viscous (ignoring the small elastic component). The effect of confining pressure is particularly evident at elevated temperatures, where fracturing is increasingly suppressed (Figure 3b). When we compare common rock types, the role of confining pressure varies considerably; for example, the effect is much more pronounced in sandstone and shale than in quartzite and slate. Thus, it appears that larger strains can be achieved with increasing depth in the Earth, where we find higher lithostatic pressures.

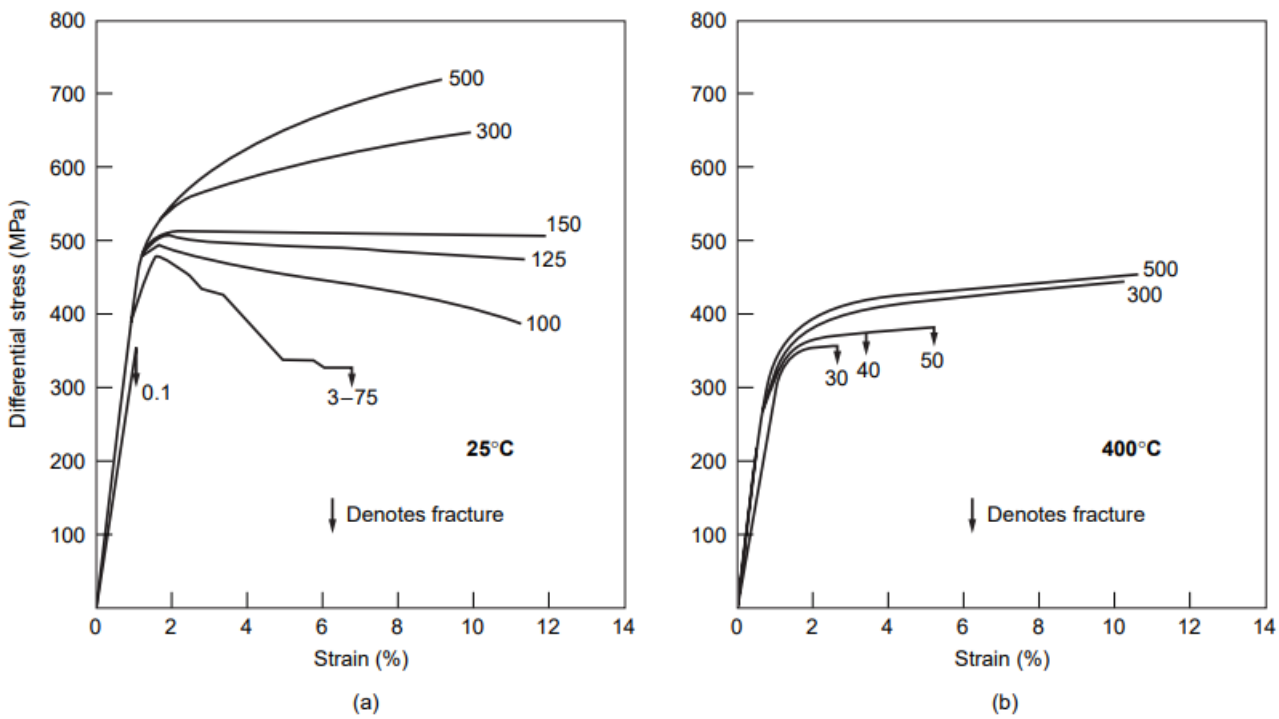

Figure 3 Compression stress–strain curves of limestone at various confining pressures (indicated in MPa) at (a) 25°C and (b) 400°C.

Temperature

A change in temperature conditions produces a marked change in behavior (Figure 4). Using the same limestone as in the confining pressure experiments, we find that the material fails rapidly at low temperatures. Moreover, under these conditions most of the strain prior to failure is recoverable (elastic). When we increase the temperature, the elastic portion of the strain decreases, while the ductility increases, which is most noticeable at elevated confining pressures (Figure 4b). You experience this temperature dependence of flow also if you pour syrup on pancakes in a tent in the Arctic or in the Sahara: the ability of syrup to flow increases with temperature. Furthermore, the maximum stress that a rock can support until it flows (called the yield strength of a material) decreases with increasing temperature. The behavior of various rock types and minerals under conditions of increasing temperature, we see that calcite-bearing rocks are much more affected than, quartz-bearing rocks. Collectively these experiments demonstrate that rocks have lower strength and are more ductile with increasing depth in the Earth, where we find higher temperatures.

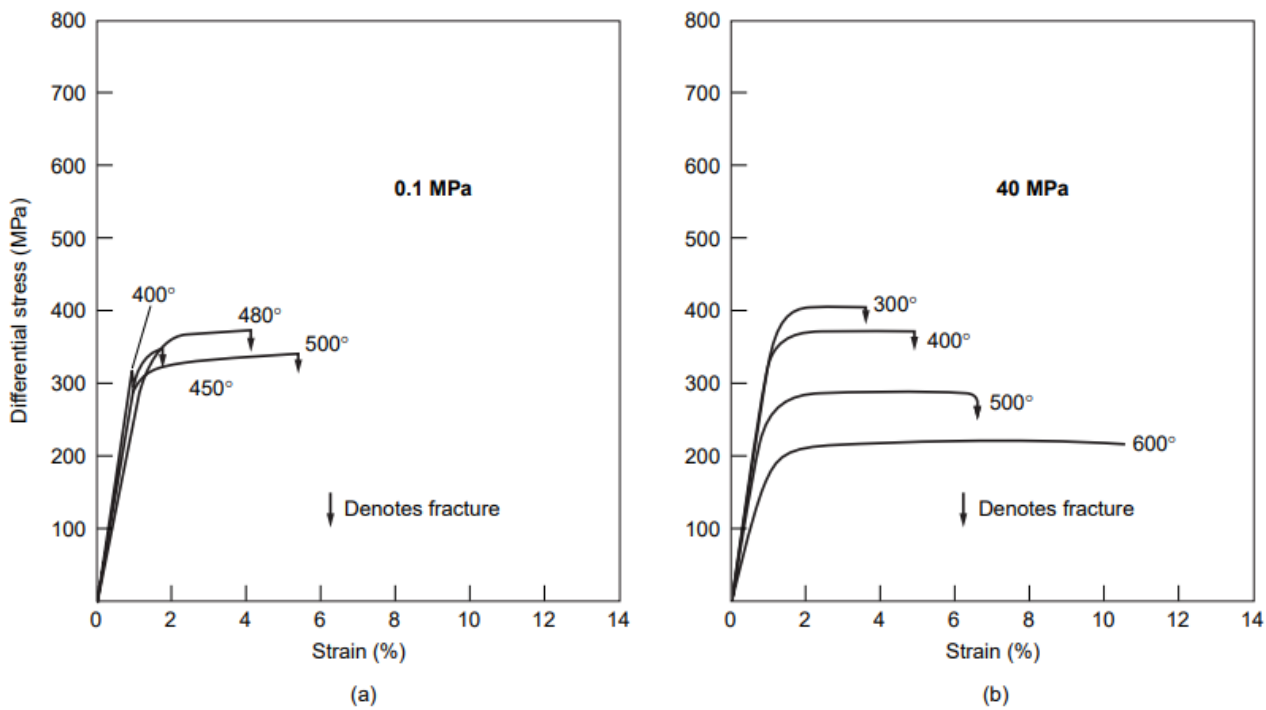

Figure 4 Compression stress–strain curves of limestone at various temperatures (indicated in °C) at (a) 0.1 MPa confining pressure and at (b) 40 MPa confining pressure.

Strain Rate

It is impossible to carry out rock deformation experiments at geologic rates, so it is particularly important for the interpretation of experimental results to understand the role of **strain rate**. The effect is again best seen in experiments at elevated temperatures, such as those on marble (Figure 5). Decreasing the strain rate results in decreased rock strength and increased ductility. We again turn to an analogy for our understanding. If you slowly press on a small ball of Silly Putty, it spreads under the applied stress (ductile flow). If, on the other hand, you deform the same ball by a blow from a hammer, the material will shatter into many pieces (brittle failure). Although the environmental conditions are the same, the response is dramatically different because the strain rate differs. Because rocks show similar effects from strain rate variation, the Silly Putty experiment highlights a great uncertainty in experimental rock deformation. Extrapolating experimental results for strain rates over many orders of magnitude has significant consequences. Consider a strain rate of 10^{-14}/s and a temperature of 400°C, where ductile flow occurs at a differential stress of 20 MPa. At the same temperature, but at an experimental strain rate of 10^{-6}/s, the flow stresses are nearly an order of magnitude higher (160 MPa). Comparing this with the results of temperature experiments, you will notice that temperature change produces effects similar to strain rate variation in rock experiments (higher t ∝ lower $\dot{e}$); $\dot{e}$ has therefore been used as a substitute for geologic strain rates. In spite of the uncertainties, the volume of experimental work and our understanding of the mechanisms of ductile flow allow us to make reasonable extrapolations to rock deformation at geologic strain rates.

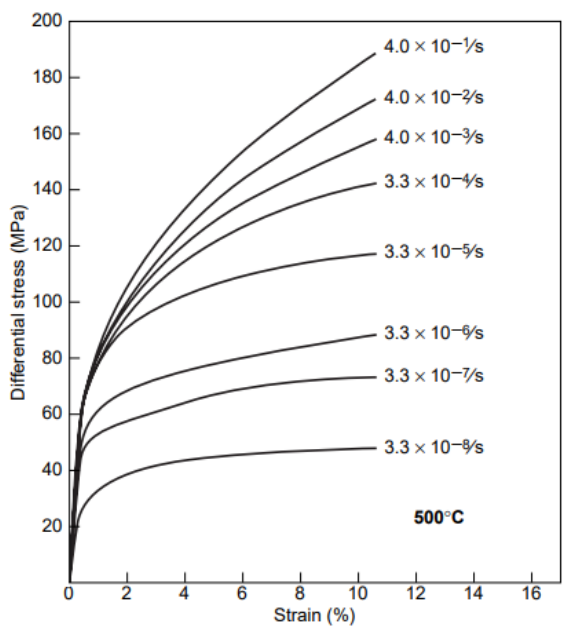

Figure 5 Stress versus strain curves for extension experiments in weakly foliated Yule marble for various constant strain rates at 500°C

Pore-Fluid Pressure

Natural rocks commonly contain a fluid phase that may originate from the depositional history or may be secondary in origin (for example, fluids released from prograde metamorphic reactions). In particular, lowgrade sedimentary rocks such as sandstones and shales, contain a significant fluid component that will affect their behavior under stress. To examine this parameter, the deformation rig shown in Figure 1 contains an impermeable jacket around the sample. Experiments show that increasing the pore-fluid pressure produces a drop in the sample's strength and reduces the ductility (Figure 6a). In other words, rocks are weaker when the pore-fluid pressure is high. Pore-fluid pressure acts equally in all directions and thus counteracts the confining pressure, resulting in an effective pressure ($P_e = P_c - P_f$) that is less than the confining pressure. Thus, we can hypothesize that increasing the pore-fluid pressure has the same effect as decreasing the confining pressure of the experiment.

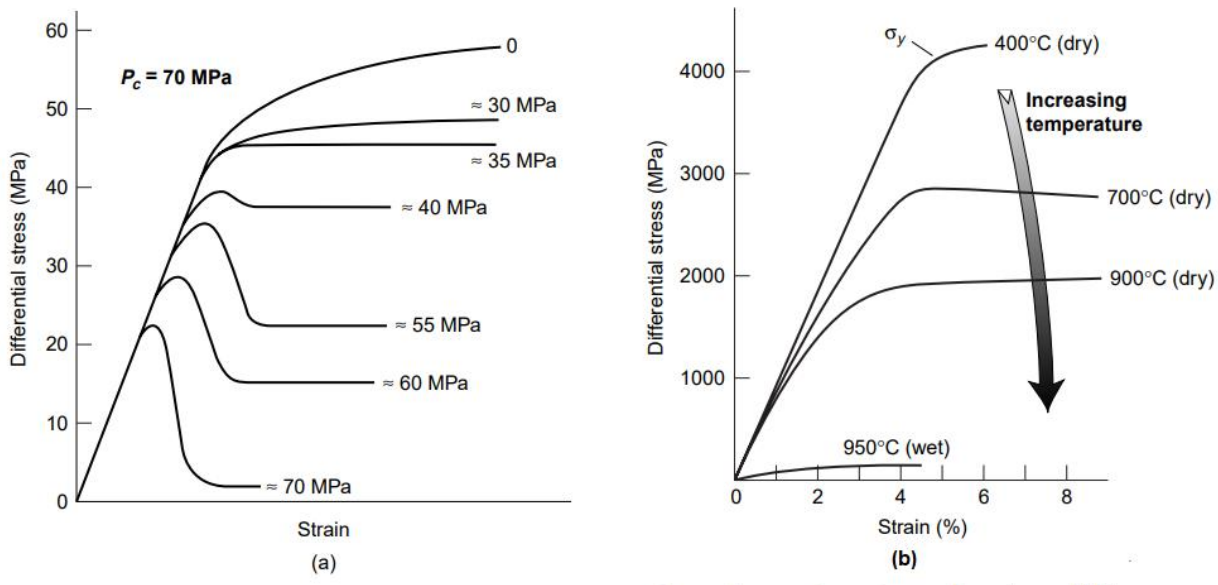

Figure 6 Comparing the effect on the behavior of limestone of (a) varying pore-fluid pressure and (b) The effect of water content on the behavior of natural quartz. Dry and wet refer to low and high-water content, respectively. The curves also show the effect of temperature on single crystal deformation, which is similar to that for rocks

The role of fluid content is a little more complex than is immediately apparent from these experiments because of fluid-chemical effects. While ductility decreases with increasing pore-fluid pressure, the corresponding decreased strength of the material will actually promote flow. The same material with low fluid content ("dry" conditions) would resist deformation, but at high fluid content ("wet" conditions) flow occurs readily. This is nicely illustrated by looking at the deformation of quartz with varying water content (Figure 6b). The behavior of quartz is similar to that in the previous rock experiment: the strength of "wet" quartz is only about one

tenth that of “dry” quartz at the same temperature. The reason for this weakening lies in the substitution of OH groups for O in the silicate crystal lattice, which strains and weakens the Si-O atomic bonds. In practice, fluid content explains why many minerals and rocks deform relatively easily even under moderate stress conditions.

References

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