Chapter 7: Ductile Creep

THE DEFORMATION of crystalline rocks involves the rearrangement of the constituent crystals by so-called *creep mechanisms*. A particular rock may flow under various physical regimes, each controlled by a particular creep mechanism and resulting in different creep laws. It is, therefore, important to distinguish the various mechanisms of creep, and these are outlined below. Distortion by crystalline creep is *permanent*, because the rock will remain distorted upon stress removal. The principal aim of this chapter is to emphasize that the interaction of the various types of crystal motions gives rise to the global or macroscopic flow, which resembles the flow of liquids. Creep laws and the generalized rheology of the lithosphere, whether ductile or in the brittle regime, are discussed in chapter eight. A detailed discussion of the subject of microstructural geology is beyond the scope of this text.

Contents: Section 7-1 explains how rock flow is accommodated on the scale of rock grains. The associated concept of steady-state foliation is clarified in section 7-2. The various crystalline creep mechanisms are systematically discussed in sections 7-3 to 7-5. Hot creep tests in the ductile regime are outlined in section 7-6. The use of the inferred deformation maps for mineral flow is explained in section 7-7.

***Practical hint:* Crystalline creep can be visualized in time-lapse videos of synthetic compounds, such as paraffin wax, paradichlorobenzene, camphor, octachloropropane, and ice. View some of the video runs made from such experiments, or, still better, conduct your own crystalline deformation experiment, using a shear device mounted on an optical microscope.**

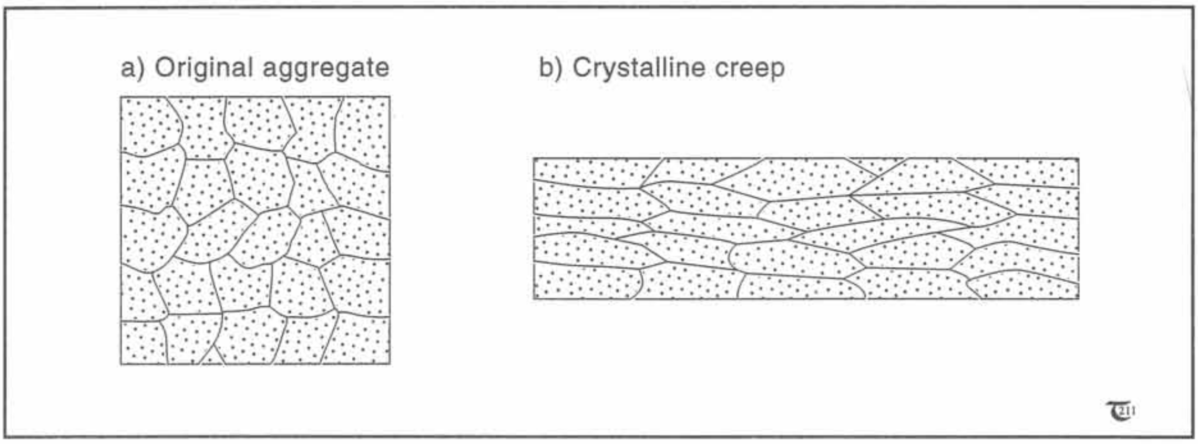


Figure 7-1: a) & b) Sketch of crystal arrangement in solid-state: (a) before deformation, and (b) after deformation by crystalline creep.

7-1 Creep of rocks

Rock is an aggregate of one or more varieties of mineral crystals and, therefore, has a range of compositions and textures. In addition to the variable composition and texture, rocks deform under a range of physical conditions. Shallow rocks are cold and subjected to low geostatic pressures, conditions which favor elastic distortion over ductile creep. Fracture and faulting occurs when the elastic limit is exceeded. Deep rocks are hot and subjected to high geostatic pressures, conditions which favor crystalline creep (Fig. 7-1a & b). Additionally, rock composition and texture, even if homogeneous, are not intrinsic properties and can both change considerably in the course of the deformation, occurring over vast periods of geological time. The texture may be modified in the course of deformation by volume changes, associated with compaction, pressure solution, diagenesis, and metamorphism. The composition may coevally alter by cementation, porphyroblastesis, pore fluid migration, fluid inclusion migration, cementing, and dia-

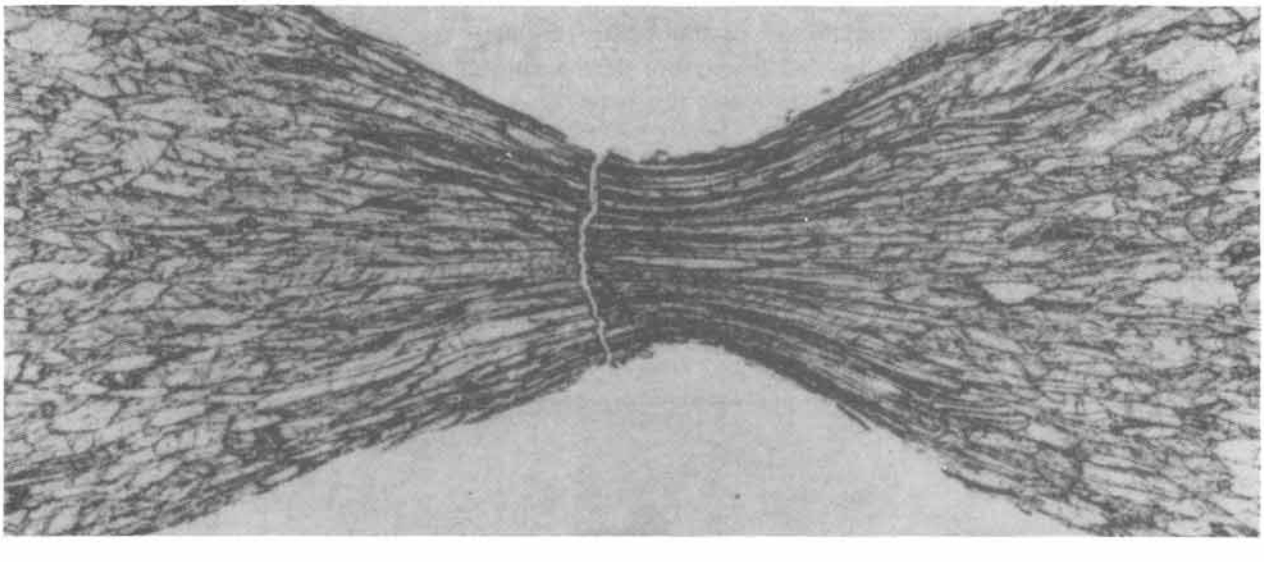


Figure 7-2a: Grain shape fabric, formed by crystalline creep in necked rod of Yule marble, extended at 300 MPa confining pressure and 600° C. Micrograph of thin section.

genesis. Any fluid pressure may greatly enhance the creep processes and associated strain-rates. Large deformation is likely to transform any isotropic rock into an anisotropic rock by cleavage development, transposition, grain-shape fabric formation, and mylonitization (Fig. 7-2a to c). The grain-shape fabric forms by crystalline creep in zones of large deviatoric stress.

Unfortunately, the way in which ductile flow in crystalline rocks is accommodated on the scale of the grains and the mode of fabric development cannot be observed in real-time. The movement of crystalline deformation or creep in rocks is extremely slow because elongation rates in active deformation zones typically are approximately 0.315 per Ma (or 10^{-14} s^{-1}). Experimental, ductile deformation of rock is possible but only under elevated temperatures and pressures for which the samples are jacketed in metal, thus, also, obstructing any real-time observation of the crystalline flow. However, some synthetic crystalline aggregates or "rocks" appear to deform by crystalline creep at room conditions and at relatively rapid rates. They can be deformed without complex experimental arrangements, and in-situ observations of crystalline flow have thus been made, using a simple shear device, mounted under an optical microscope. Materials used in such studies are the following: camphor (analog for feldspar), paradichlorobenzene (analog for mica and feldspar), and octachloropropane (analog for quartz and olivine).

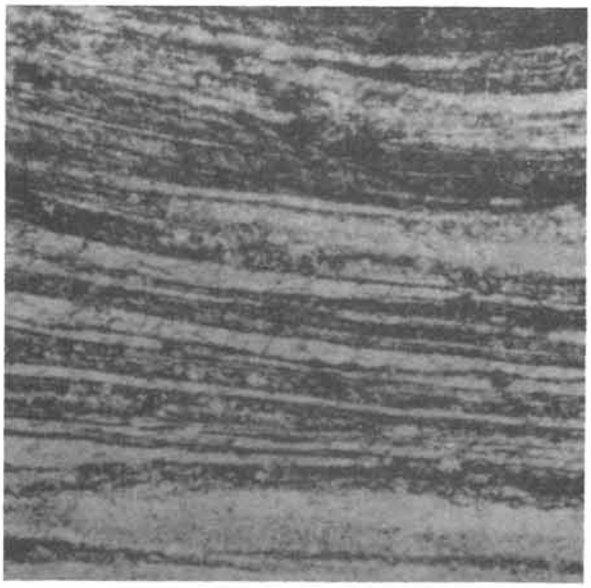


***Figure 7-2b:** Detailed view of the neck region in the Yule marble rod of Figure 7-2a.*

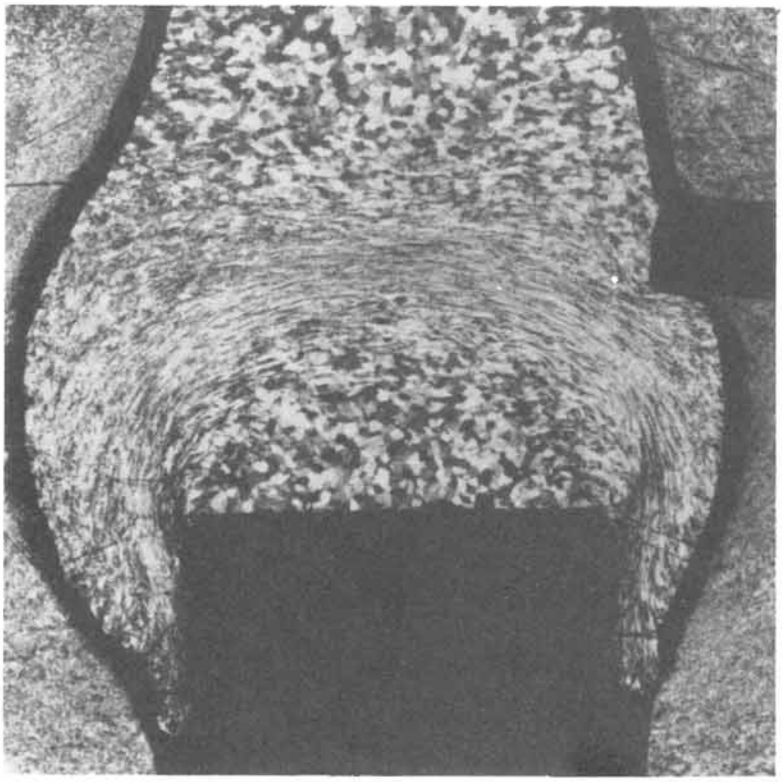
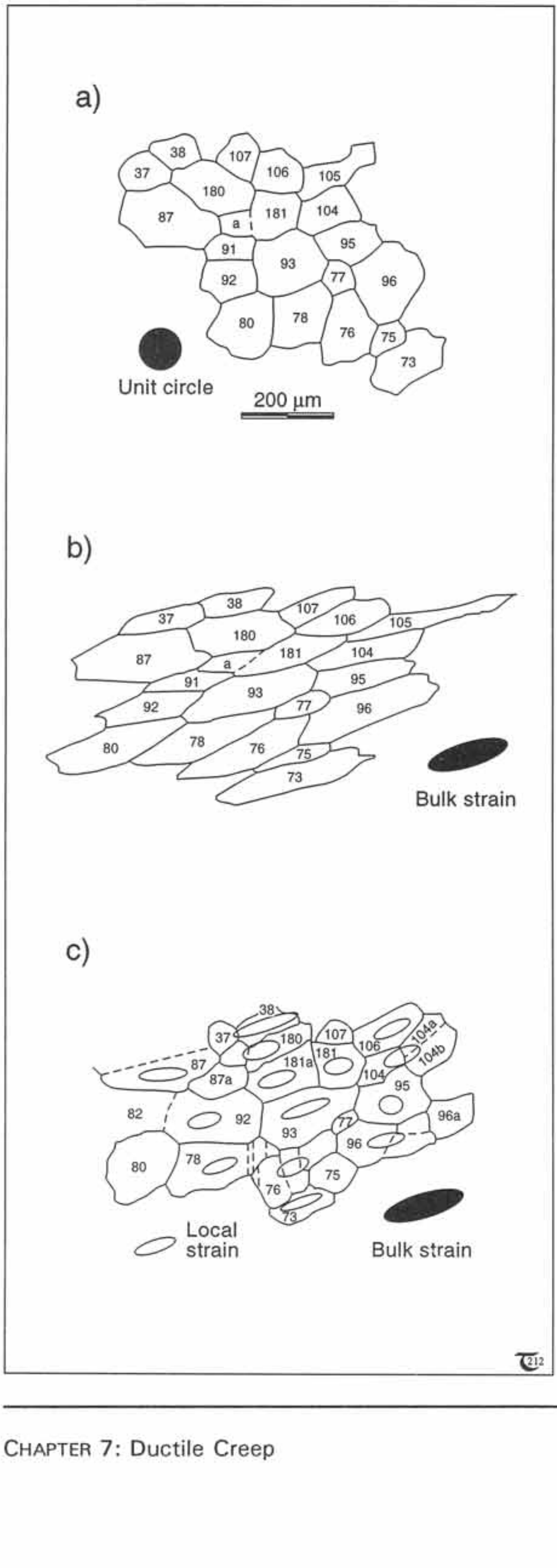


***Figure 7-2c:** Quartzite mylonite fabric around a square indenter, pushed into the rock sample. Micrograph of thin section.*

Time-lapse videotapes of the experimental deformation, now widely available (see references herein), demonstrate how such synthetic compounds accommodate crystalline flow. Initially straight grain boundaries become gradually lobated and then migrate and bifurcate to form subgrains. The formation of these *subgrains*, new areas where the crystal lattice locally mismatches that of the initial parent grain, due to activation of interatomic processes, eventually effectuates a reduction in grain size. If the mismatch in crystallographic orientation of adjacent grains becomes larger than about 7°, their contact surface will have become sufficiently mismatched so as to be regarded as a new crystal boundary. Small, new crystals continually grow at the expense of other, older crystals, and this renewal of grain shape explains the macroscopic flow of rock. The continuous process of grain modification in solid-state of crystalline aggregates in response to externally applied stresses is termed *dynamic recrystallization*.



□Exercise 7-1: Watch a videotape, showing dynamic recrystallization, and subsequently summarize in a brief essay what has been shown.

7-2 Steady-state foliation

Synkinematic microscopy of transparent polycrystals has, also, revealed that dynamic recrystallization creates a grain-shape fabric which tends to align with the major bulk strain axis. However, the grain boundaries of crystals do not deform as passive strain markers. Figures 7-3a to c illustrate the initial texture of an undeformed crystalline aggregate (a), predict what the grain boundaries would have looked like if responding as passive markers to the applied bulk strain (b), and compare this with the actual grain texture observed after the same bulk strain (c). The actual grain fabric is less deformed than expected on the basis of the bulk strain (Figs. 7-3b & c). It can be concluded, therefore, that the grain shape has a limited *strain memory*. Additionally, the externally applied bulk strain is accommodated by the grains in highly inhomogeneous fashion.

Figure 7-3: *a) to c) Experimental study of fabric development by dynamic recrystallization. (a) Undeformed fabric with labelled grains. (b) Predicted fabric for deformation of massive grain boundaries by bulk strain indicated. (c) Actual fabric, observed after deformation.*

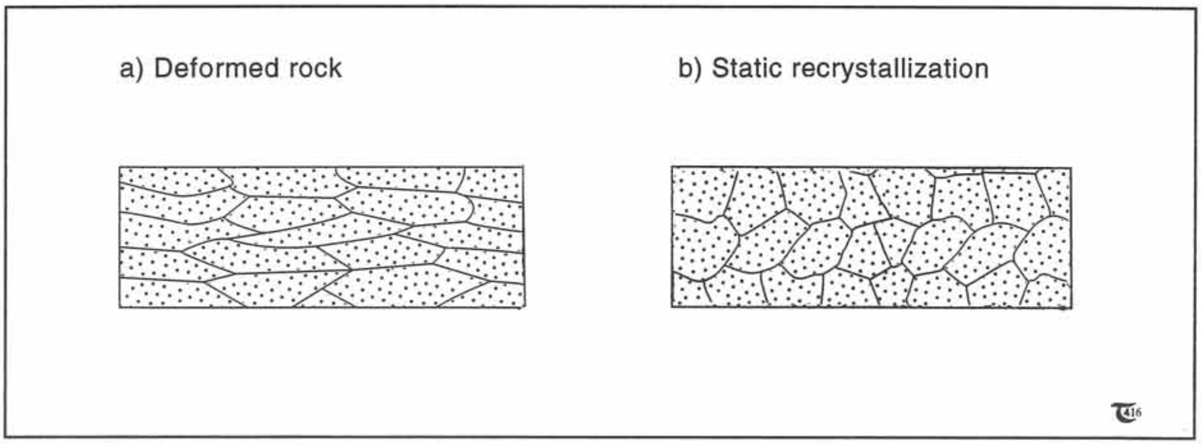


Figure 7-4: *a) & b) Grain shape fabric, due to dynamic recrystallization (a), may return to polygonal pre-stressed state by static recrystallization (b), if conditions are favorable.*

One possible explanation for the poor strain memory of crystal grains inside rocks is that strained crystal grains tend to attempt, very slowly, to regain their undistorted polygonal shape if the deformation ceases or if stress is suddenly removed (Figs. 7-4a & b). This process, termed *static recrystallization*, is attributed to recovery of the low-energy state of crystals. Static recrystallization is particularly successful in laboratory tests, after removal of the externally applied stresses, while maintaining high temperatures.

Even during deformation, the process of strain reduction is competing with the processes of dynamic distortion, and their rates are likely to balance at some stage, a process termed *steady-state foliation*. The grain-shape fabric will display a strain, lower than that of the actual bulk strain. As deformation continues, the bulk strain will progressively increase, but this increase is not reflected in the grain-shape fabric of a steady-state foliation. Grain-shape fabrics in natural rocks can be preserved, once rocks are brought to the Earth's surface by geodynamic processes. This is because static recrystallization cannot occur in most rocks under the physical conditions prevailing at the surface.

□Exercise 7-2: Explain: a) What is meant by strain memory of grain-shape fabrics? b) Why does ductile creep cause irreversible macroscopic deformation, even if static recrystallization were to occur?

□Exercise 7-3: Consider a steady-state grain-shape fabric. If the strain-rate decreases to zero, static recrystallization is likely to obliterate the preferred orientation of grain-shape fabrics. On the other hand, do you expect differences in the intensity of grain-shape fabrics at different steady-state strain-rates? Elaborate upon your answer.

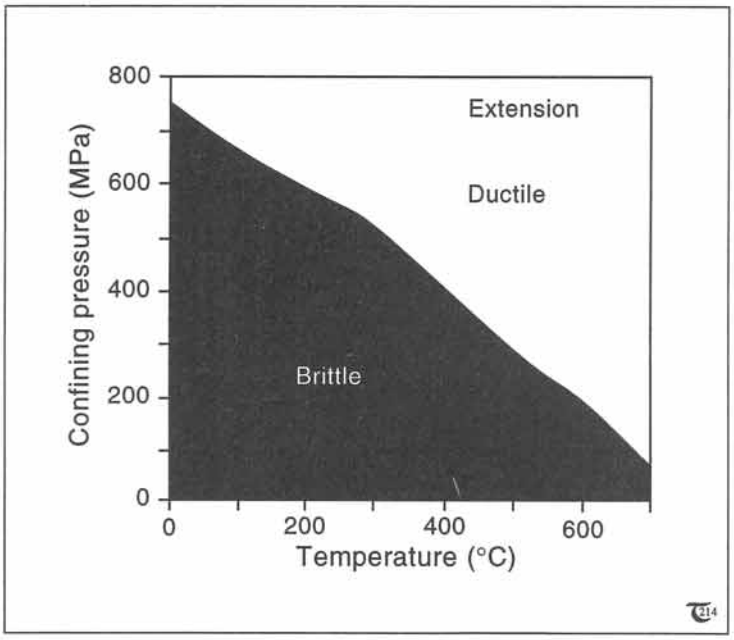


Figure 7-5: Generalized diagram of the brittle-ductile transition observed in creep tests over a range of pressures and temperatures.

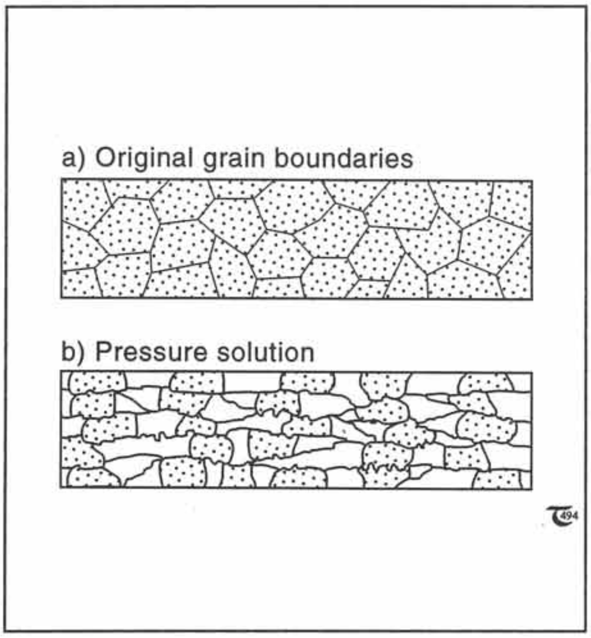


Figure 7-6: a) & b) Deformation of rock volume by pressure solution.

7-3 Creep mechanisms

The ductile deformation of crystalline rocks may be controlled by a range of microscopic mechanisms (occurring on scales from ions to mineral grains). The principal mechanisms are: *pressure solution*, *diffusion creep*, *dislocation creep*, and *superplasticity*. These four mechanisms are determined by the physical conditions, prevailing when the deformation takes place, and the major controlling factors are pressure, temperature, fluid pressure, and strain-rate. Creep tests with the triaxial Griggs apparatus show that rocks will generally break if deformed at surface temperatures, unless the confining pressure is raised to extremely high values of over 700 MPa (Fig. 7-5). However, ductile creep may occur at much lower pressures if the temperature is corresponding to those at larger crustal depths. The so-called homologous temperature, $T_H = T/T_{melt}$, plays a crucial role in whether or not crystalline creep can occur. The closer the homologous temperature gets to unity, the easier it is for crystals to creep in a ductile fashion. Actual melting and disintegration of the crystal lattice occurs when T_H reaches unity, for which crystalline creep is no longer defined.

Pressure solution or *solution transfer* is a creep mechanism, which is basically different from the others, because it is largely controlled by chemical erosion of the grain boundaries, assisted by hydrothermal circulation. Material is forced into solution at points of high pressure and is precipitated in regions of low pressure. It results in a grain-shape fabric without any trace of deformation in the grain interior and is accompanied by significant volume loss (Figs. 7-6a & b). Pressure solution may operate at near surface conditions, where rocks would normally fault, and is likely to play an important role in the development of cleavage fabrics in slates and shales. Constitutive relationships for solution transfer are not available, as it operates in an open chemical system,

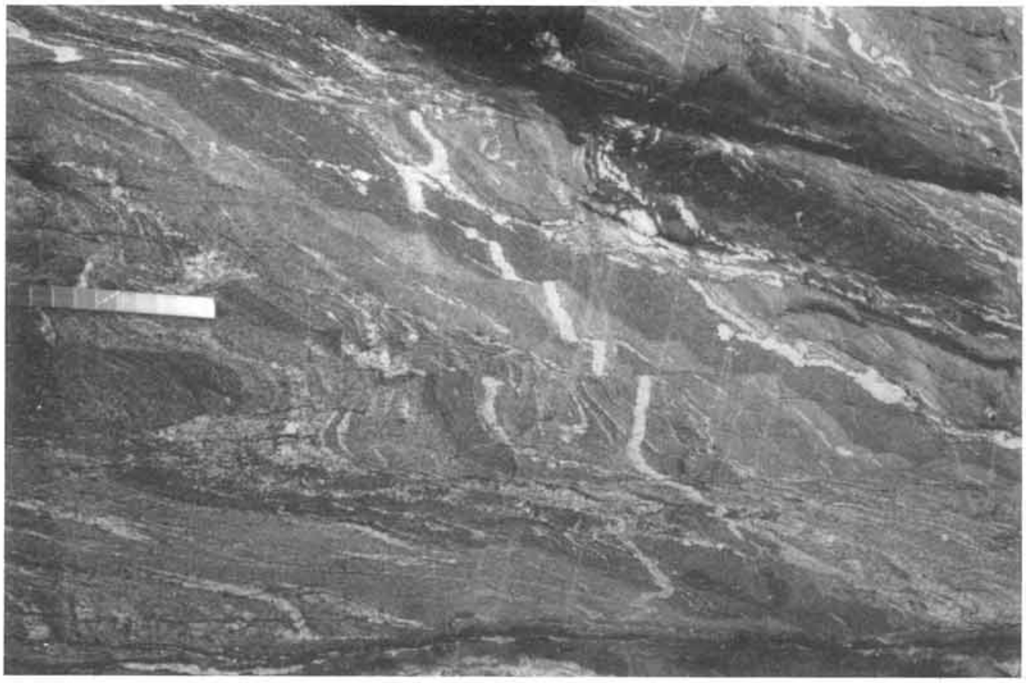


Figure 7-7a: *Horizontal pressure solution seams in Precambrian gneiss of Scandinavian Shield, Vaermlandsnaes, Sweden.*

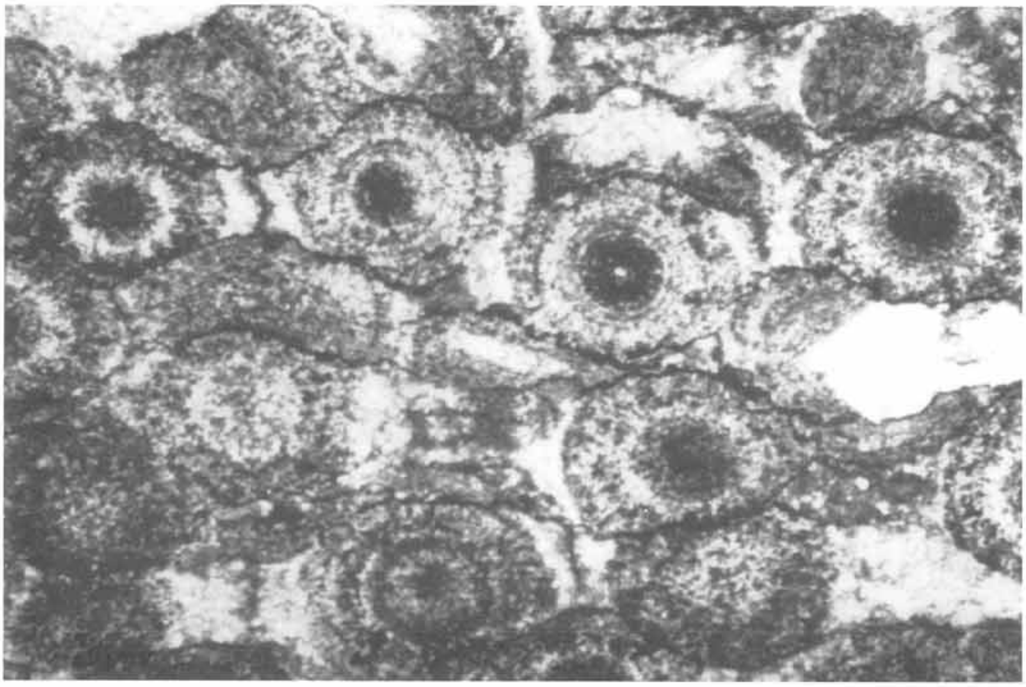
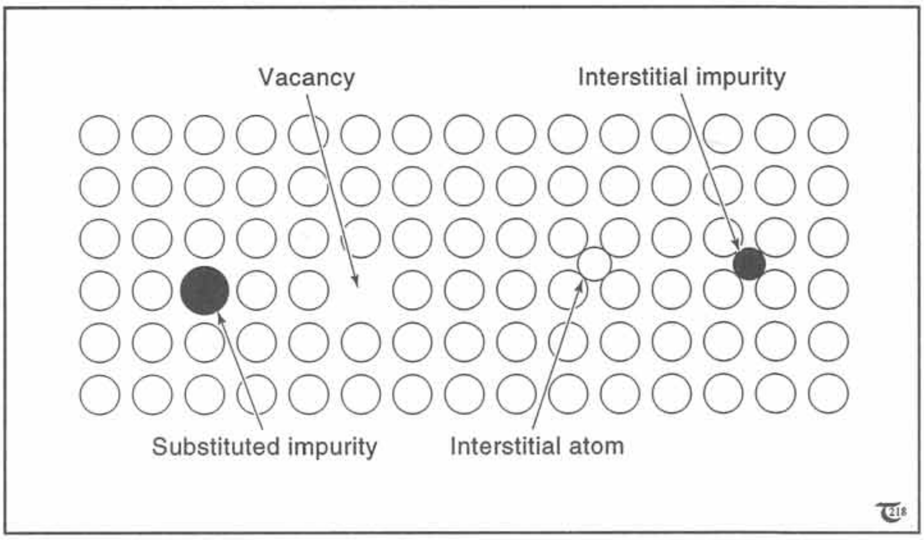


Figure 7-7b: *Thin section of stylolite seams eroding ooids of oolitic limestone, accommodating vertical shortening of the rock volume.*

Figure 7-8: Various possible point imperfections of the atomic arrangement in a crystal lattice.



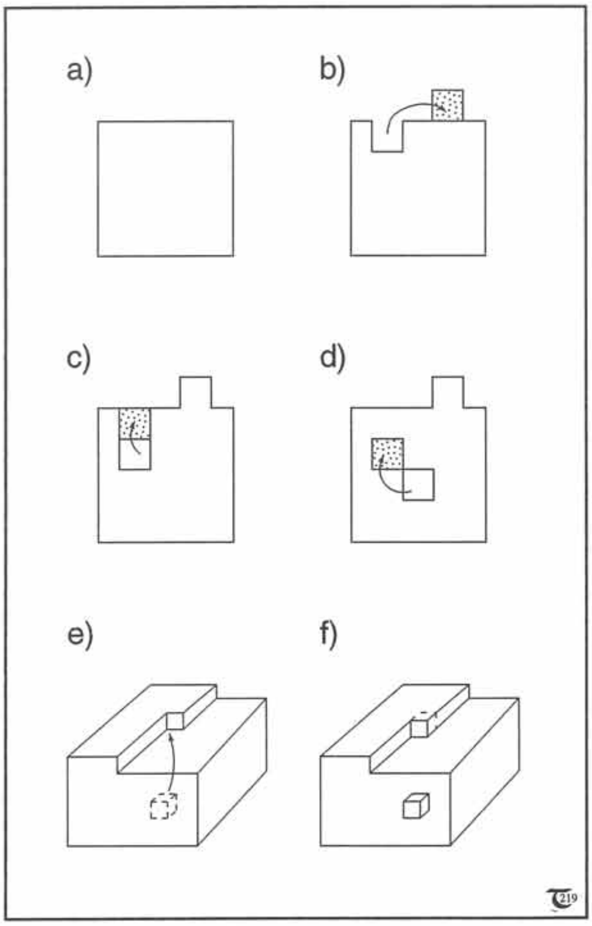


Figure 7-9: a) to f) Schematic representation of the creation and movement of vacancies in a crystal lattice to achieve changes in the external shape of mineral grains.

involving chemical potential of the circulating fluids, pore space reduction, and bulk volume change. It is, in fact, unlikely that a constitutive relationship between stress and strain-rate can ever be established, as this would require steady-state solution transfer, which seems improbable. Pressure solution is the only creep mechanism which may operate in the absence of a tectonically induced deviatoric stress, and the *stylolites* or serrated solution seams, resulting from local solution transfer, are commonly found perpendicular to the directon of lithostatic loading (Figs. 7-7a & b). Insoluble heavy minerals often are concentrated along such pressure solution seams.

□Exercise 7-4: Stylolites or pressure-solution seams are good features to constrain the orientation of the principal paleostresses. Explain.

7-4 Diffusion creep

The crystal lattice of the mineral grains in solid rock, a regular assembly of one or various atoms, usually contains imperfections or lattice defects. Three common types of *point imperfections* are: empty lattice positions or vacancies, lattice positions filled by impurities, and interstitial atoms (either foreign or from the compound itself, i.e., self-interstitial) (Fig. 7-8). Diffusion or movement of any atom through a crystal lattice may occur by mobilizing the point imperfections, for example, by serial jumping of atoms into neighboring vacancies. Any deviatoric stress may induce gradients in the vacancy concentration of a crystal, provided the *activation temperature* is reached. Vacancies may then migrate along the gradient to enable displacement of atoms in the opposite direction, which is macroscopically expressed as a change in the crystal shape. New vacancies may be generated at the grain boundaries (Figs. 7-9a & b).

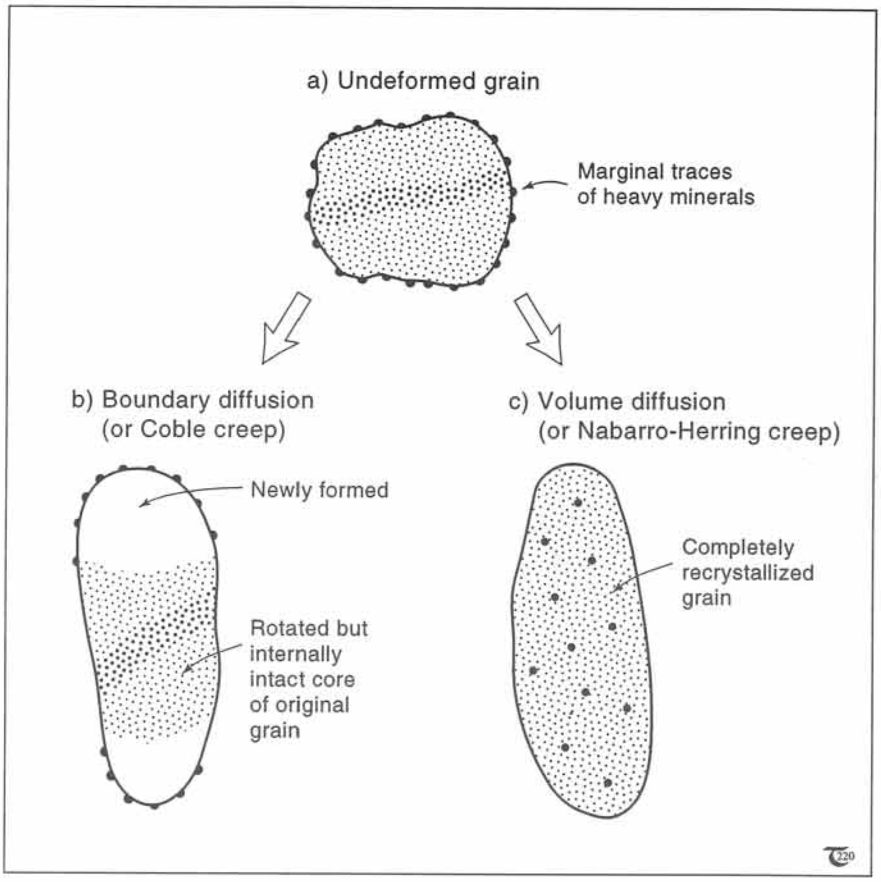


Figure 7-10: *a) to c) Optical effects of two main types of diffusion on the appearance of a hypothetical grain.*

Two major modes of diffusion are distinguished: grain boundary diffusion or *Coble creep* and volume diffusion or *Nabarro-Herring creep* (Figs. 7-10a to c). Both these mechanisms are linear creep processes, that is, the ratio of the stress and strain-rate is constant for any stress values. However, the viscosity, a flow parameter discussed in more detail in chapter eight, is proportional to the exponent of the temperature. In addition, the viscosity for minerals in Coble creep is, also, proportional to the third power of the grain size and in Nabarro-Herring creep to the second power of the grain size. Grain size reduction, during dynamic recrystallization, therefore, will significantly decrease the effective viscosity of the rock and may lead to concentration of the flow in regions of lowered viscosity.

Diffusion creep is generally *not* observed in rock samples deformed under laboratory conditions, because dislocation creep is likely to dominate under the relatively high strain-rates used, i.e., between 10^{-9} and 10^{-2} s^{-1}. Measurements at strain-rates lower than 10^{-9} s^{-1} would imply experiments over impractically long periods of years, decennia, centuries, millenia, etc. However, it can be calculated on the basis of theoretical grounds that diffusion or linear creep is more energy-efficient at the low strain-rates of 10^{-14} s^{-1} of most natural deformations of common rock types and, therefore, is abundantly operating in nature. But, of the two mechanisms of diffusion, Nabarro-Herring creep needs very high homologous temperatures for its activation and, thus, is less common in crustal rocks than Coble creep.

□Exerc. 7-5: Why can grain-size reduction cause softening of rock, deforming by diffusion creep?

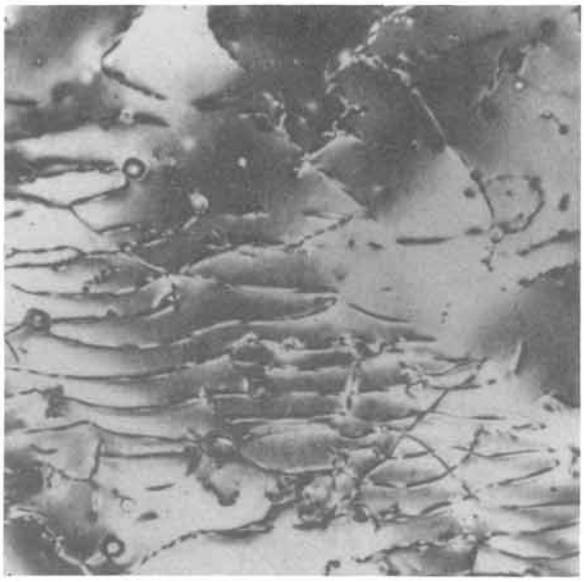


Figure 7-11a: Transmission electron micrograph of quartz.

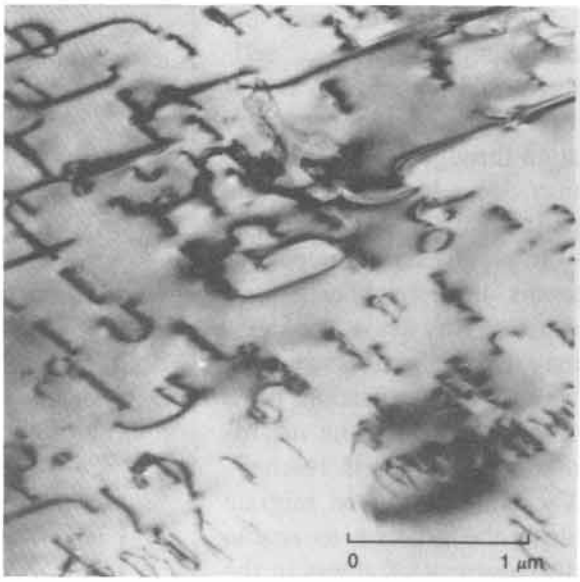


Figure 7-11b: Transmission electron micrograph of olivine.

7-5 Dislocation creep

The crystal lattice of most minerals possesses one or more planes of weak bonding, determined by the crystallographic symmetry. Such potential glide planes allow the creation and removal of dislocation planes and dislocation lines. *Dislocation glide* or *crystal plasticity* is possible at $T_H < 0.5$, whereas *climb* of dislocations out of the glide plane is possible only at $T_H > 0.5$. Dislocation creep will usually dominate over other creep mechanisms in experimental dynamic recrystallization. Transmission electron microscope studies of etched crystal surfaces after deformation tests reveal the presence of numerous dislocation traces (Figs. 7-11a & b). The internal patterns are crystalloplastic microstructures due to deformation, commonly including dislocation lines, loops, and tiny trails of fluid inclusions. Deformation by dislocation creep may cause development of anisotropic textures by crystal rotation, twinning, kink-band formation, and slip-band formation (Figs. 7-12a & b).

Dislocation glide is assisted by two main types of dislocations: *edge and screw dislocations*. Edge dislocations are aligned along the leading edge of a slip plane, propagating through a crystal lattice (Figs. 7-13a to e). The dislocation line inside the lattice is likely to be normal to the external shear direction. Screw dislocations can be visualized as similar to tearing a stack of paper. The dislocation, marked by the tear, terminates at the dislocation line, running through the sheet at the propagating tip of the tear, parallel to the shear direction (Fig. 7-14a to c). The atom planes in a crystal, associated with this type of dislocation are distorted helicodially. A glide plane may be laterally terminated by a combination of edge and screw dislocations (Fig. 7-15).

Ideally, minerals should comprise five independent slip directions to accommodate triaxial deformation of an incompressible solid without necessity of lattice rotation. In practice, many minerals possess fewer glide planes. For example, mica possesses only one good (001)-glide plane, and plastic bending is achieved by the movement of edge dislocations through the (001)-plane (Fig. 7-16a to d). The result of the plastic bending will be a gradual change in the undulous extinction along the arc of the folded mica (Fig. 7-17). Finally, such edge dislocations may pile up to form a high-angle grain boundary, effectively

Figure 7-12a: *Micrograph of quartz mylonite in shear zone beneath the Seve nappe, Scandinavian Caledonides. Stretching of the grains is principally due to dislocation creep. Courtesy Hakan Sjöström.*

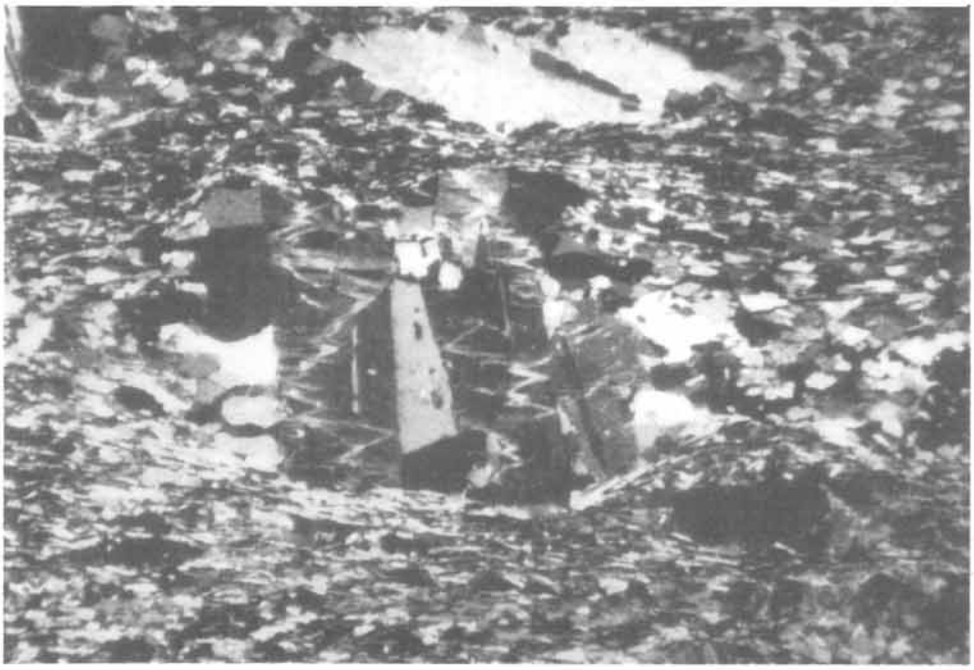
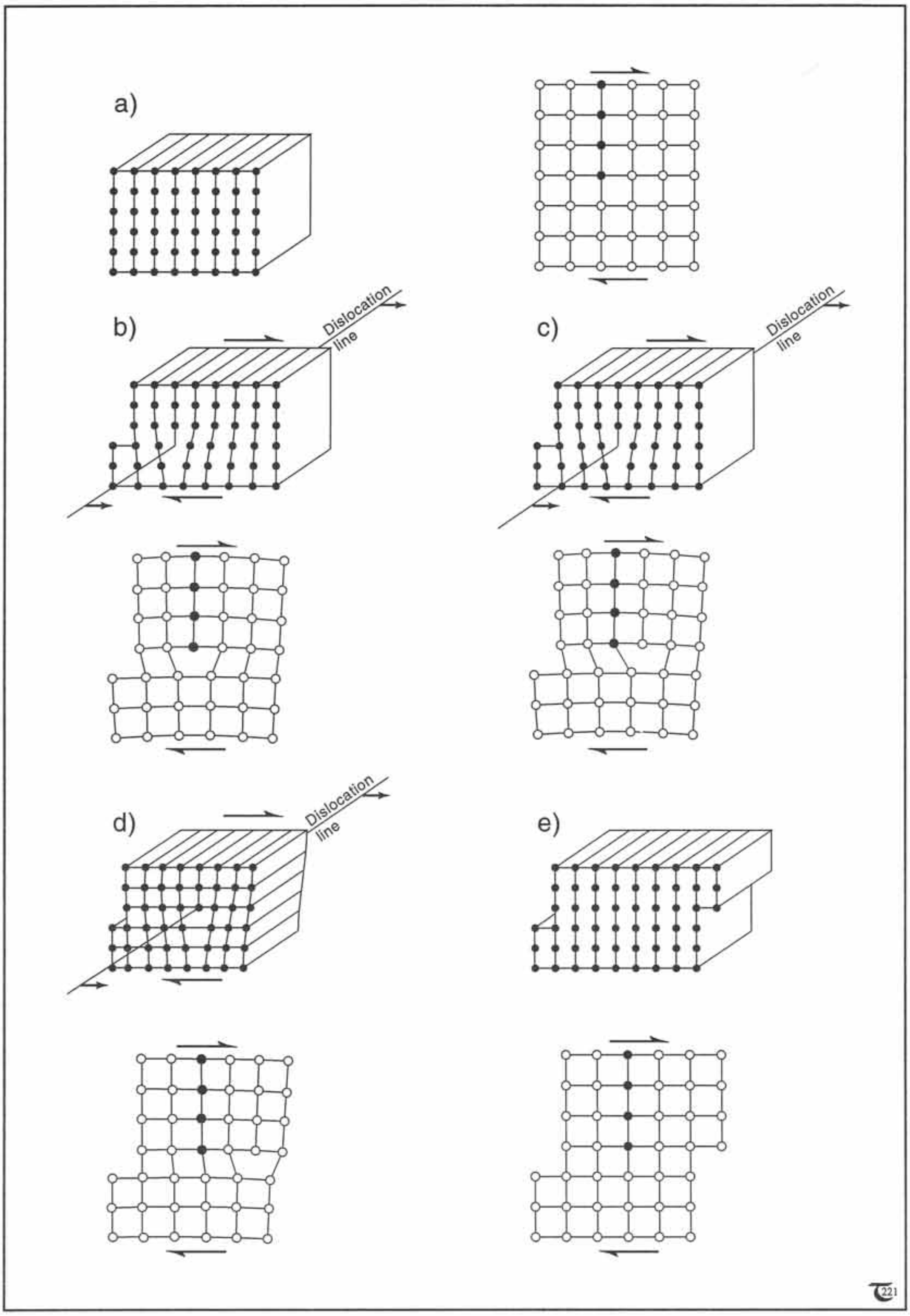


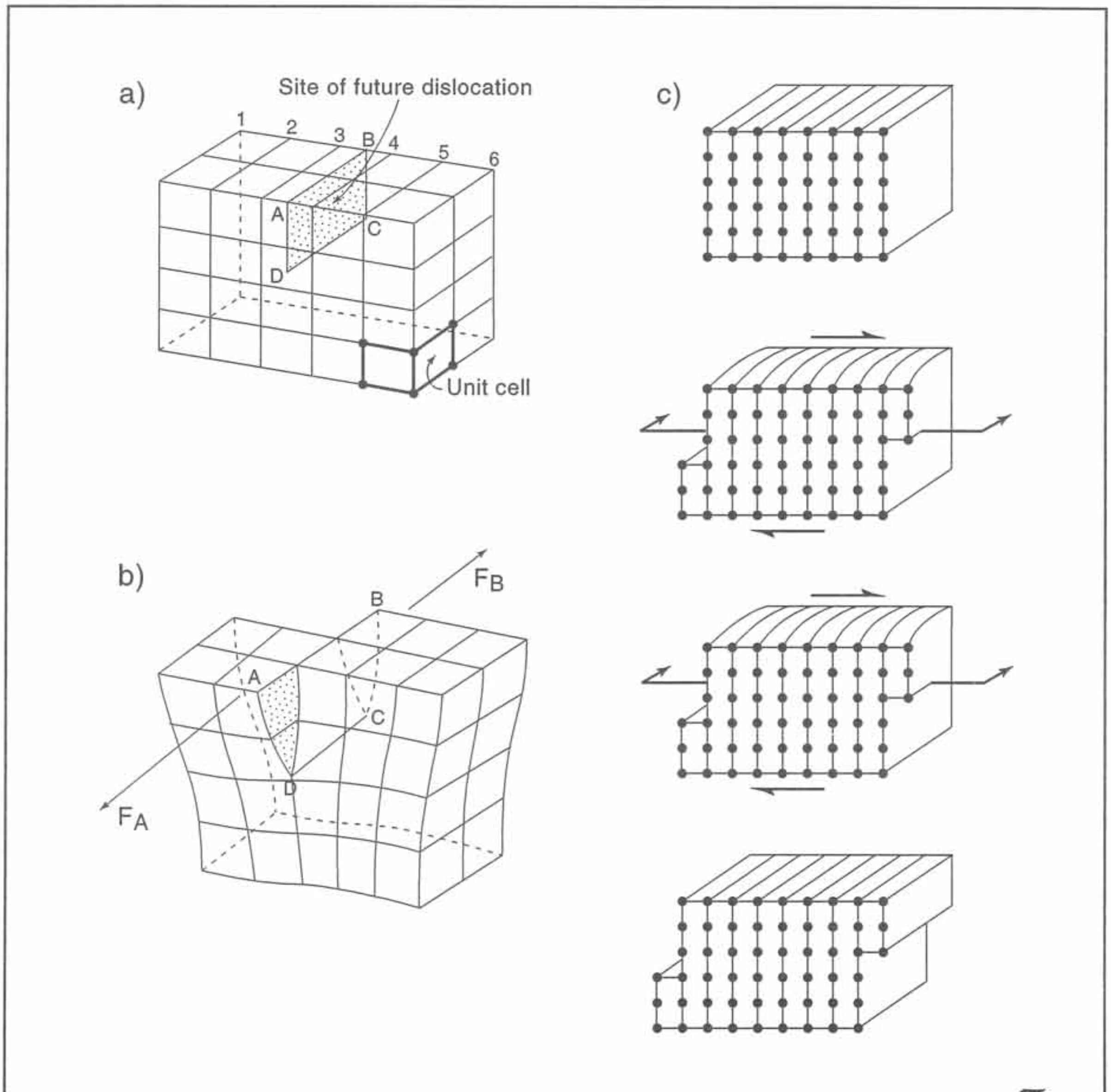
Figure 7-12b: *Micrograph of kinked biotite porphyroblasts in schist from the Seve nappe, Scandinavian Caledonides, Sweden. Kinkband formation is mainly due to dislocation creep. Courtesy Håkan Sjöström.*



a)
b)
Dislocation line
c)
Dislocation line
d)
Dislocation line
e)
221

Figure 7-13: *a) to e) Schematic sequence of the kinematics of a dislocation line, marking an array of edge dislocations, gliding through a crystal lattice, so as to accommodate macroscopic deformation of the crystal boundary.*

Figure 7-14: *a) to c) Kinematics of a screw dislocation. (a) Atomic bonds across plane ABCD are disrupted by tensing forces F_A and F_B (b). The initiation and removal of screw dislocations provides a mechanism for internal glide plane movements, changing the external shape of crystals (c).*



(Continued from p. 112) removing the undulous extinction from the mica lattice. The removal of dislocations by sweeping them into old and new grain boundaries is called *recovery* or *annealing*.

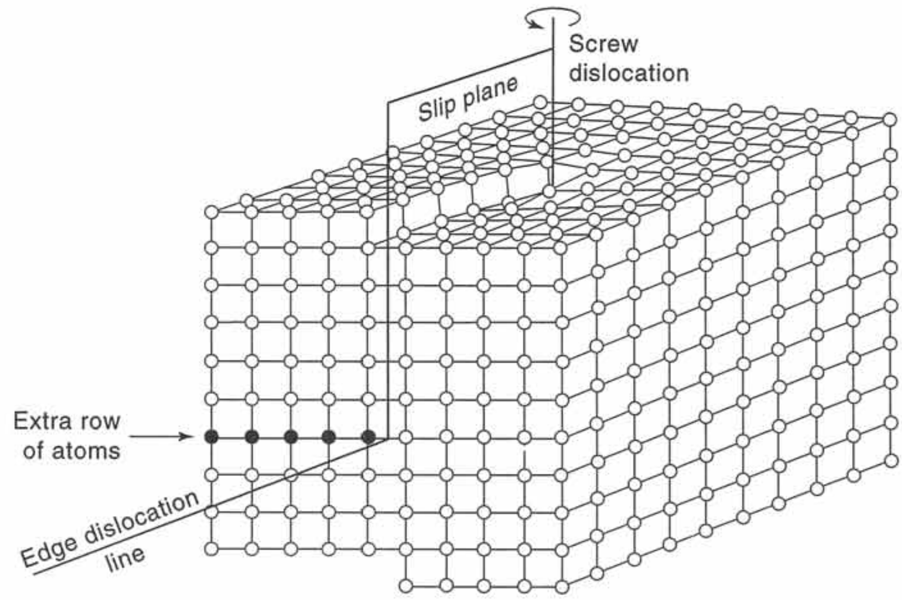


Figure 7-15: *Edge dislocations and screw dislocations may operate coevally and terminate against one another.*

□Exercise 7-6: Figure 7-18 shows the microfabric of a schist with buckled mica and undulatory extinction of both mica and quartz between the arches of mica. Explain the creep mechanism that may have been operating to assist the formation of the deformation fabric in this schist.

□Exercise 7-7: a) Explain how stress activated dislocation glide or movement of edge dislocations creates undulous extinction in an initially undistorted grain. b) Explain how subgrains may evolve into high-angle grain boundaries through the mechanism of using edge dislocations.

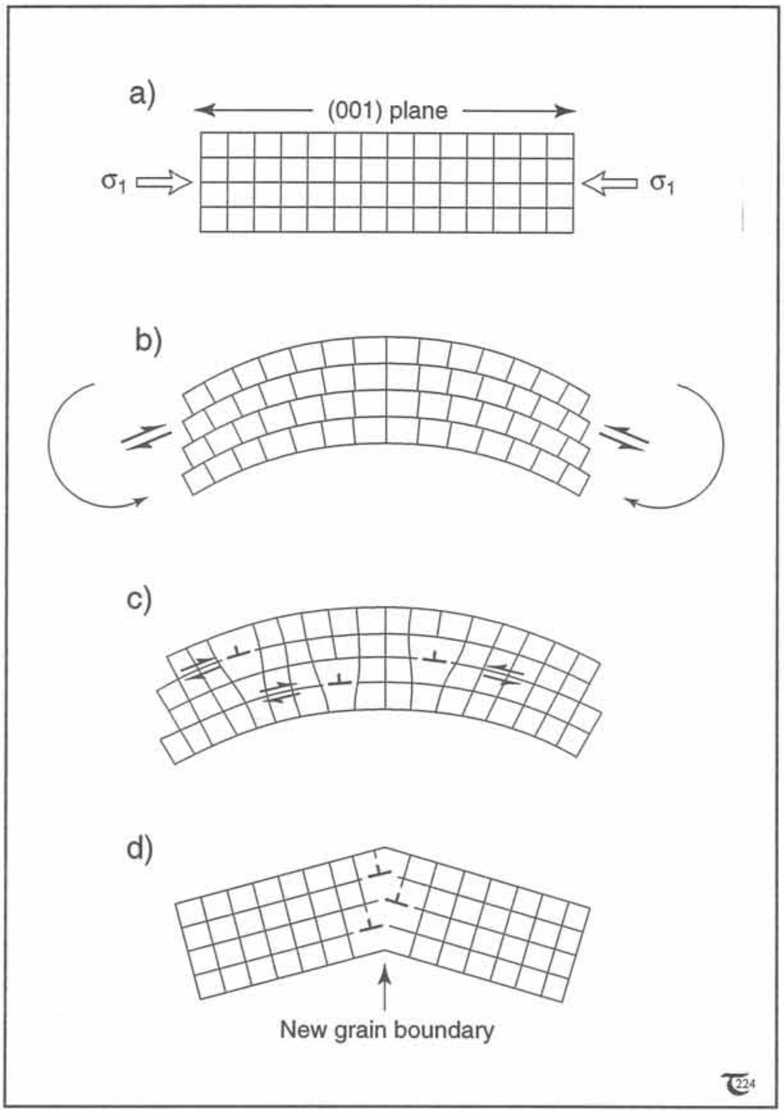


Figure 7-16: *a) to d) Sections of a mica crystal-lattice: (a) before the deformation, (b) during initial buckling, (c) after formation of edge dislocations, and (d) after formation of a new grain boundary by the piling up of dislocations.*

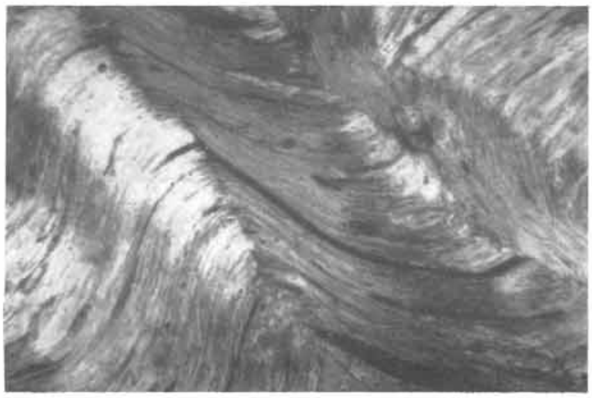


Figure 7-17: *Micrograph of undulatory extinction in buckled micas.*

Figure 7-18: *Micrograph of deformed quartz-mica schist. See exercise 7-6.*

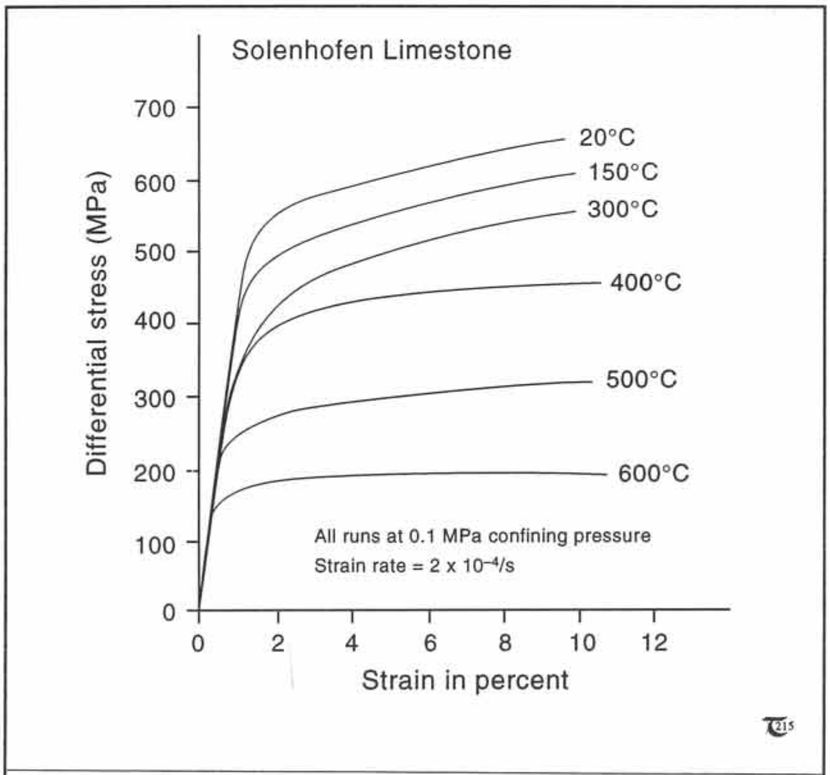


Figure 7-19: Stress-strain curves for Solenhofen Limestone, deformed at a constant strain-rate of $2x10^{-4}$ s^{-1}, atmospheric pressure, and a range of temperatures, as indicated.

7-6 Creep tests

Diffusion creep, dislocation creep, and superplasticity are the principal mechanisms through which rocks may deform in the ductile regime. The resulting mode of macroscopic flow is distinctive and can be understood better considering some results from tests in rock mechanics. Triaxial tests at elevated confining temperatures and pressures will generally record the mechanical response of crystalline rock deforming by ductile creep, rather than by fracturing, but the respective stress-strain curves look very similar in shape. Figure 7-19 illustrates a stress-strain curve for Solenhofen Limestone at atmospheric pressure and at temperatures between 20° and 600° C, using a constant strain-rate of $2x10^{-4}$ s^{-1}. The hotter test displayed creep fabrics when recovered from the test rig; the colder samples only fractured.

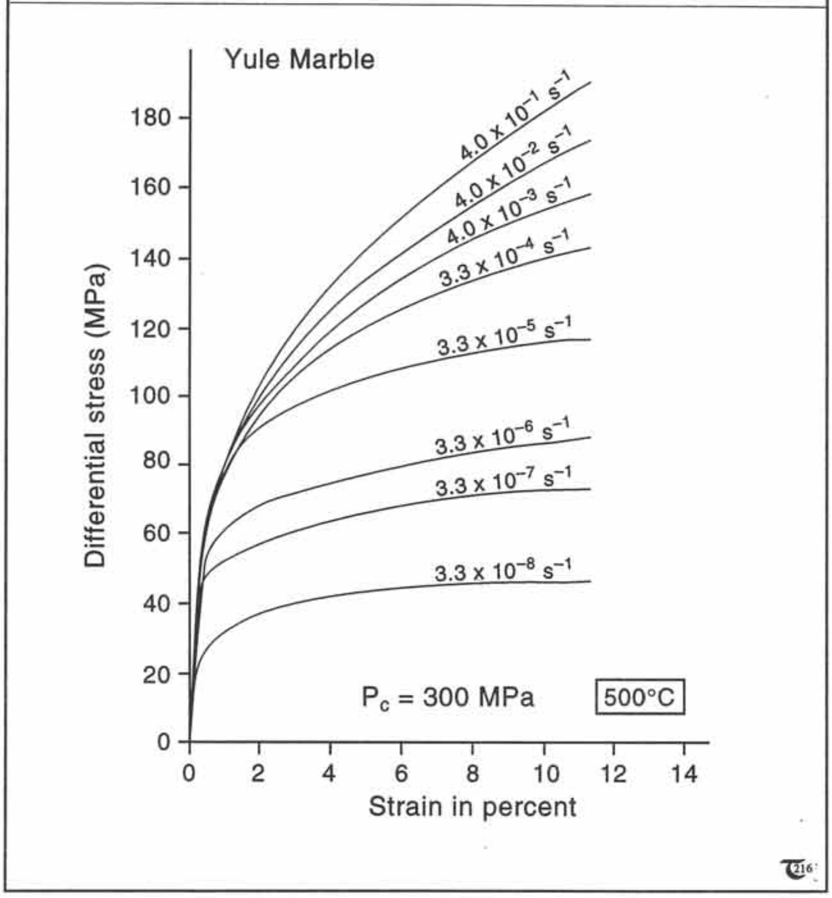
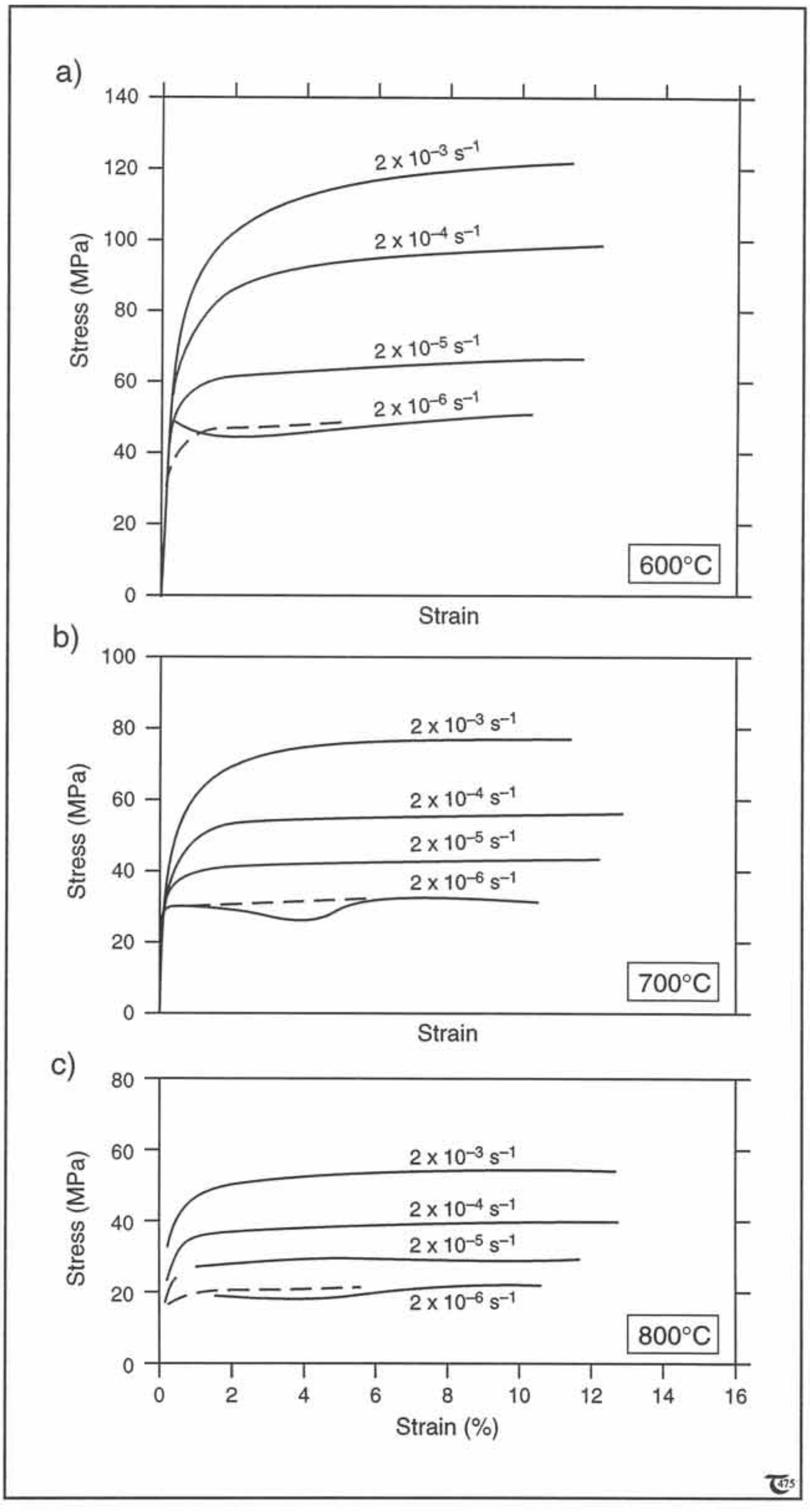


Figure 7-20: Stress-strain curves for Yule Marble, deformed at a constant temperature of 500° C, elevated pressure of 300 MPa, and a range of strain-rates, as indicated.

***Figure 7-21:** a) to c) Stress-strain curves for Yule Marble, deformed in three sets of isothermal runs (600°, 700°, and 800° C) for the range of strain rates indicated.*



In order to determine creep laws, the temperature and pressure are fixed well into the ductile creep regime. Figure 7-20 shows stress-strain curves for Yule marble at (T,P) fixed at (500° C, 300 MPa); each curve differs about an order of magnitude in the strain-rate applied. These data can be more summarily represented in *log stress-log strain-rate* plots, for which it is important that steady-state flow results are used. *Steady-state creep* is established when the stress reaches a constant plateau-value on the stress-strain graph. For any applied strain-rate there is initially an increase in the stress (primary creep), which attains a constant value (the steady-state stress) after some time. It is apparent from Figures 7-21a & b that only approximately steady state is reached.

The steady-state rheology of rocks is of great importance to establish constitutive relationships involving stress, strain-rate, and intrinsic flow parameters for the rock (see chapter eight). One particular strain-rate yields only one particular deviatoric stress in steady-state. The experimental

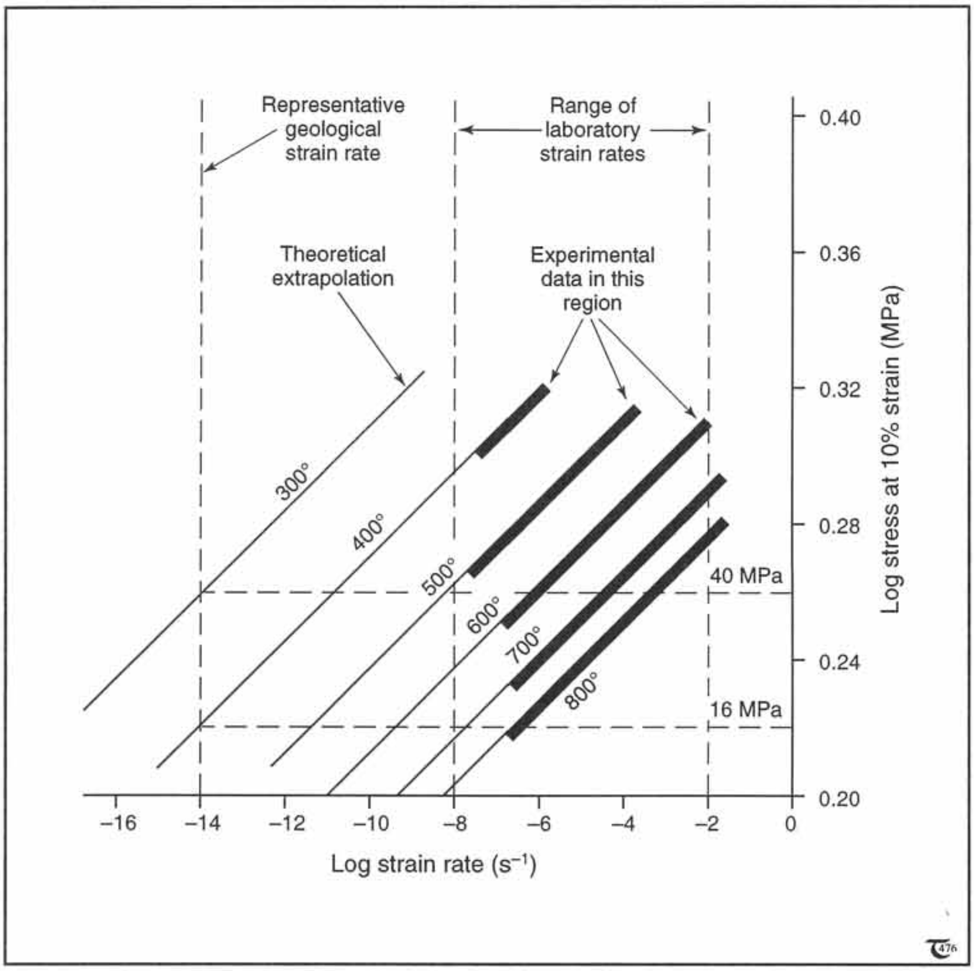


Figure 7-22: Stress-strain flow curves for Yule marble in log stress-log strain rate plot for a range of different temperatures.

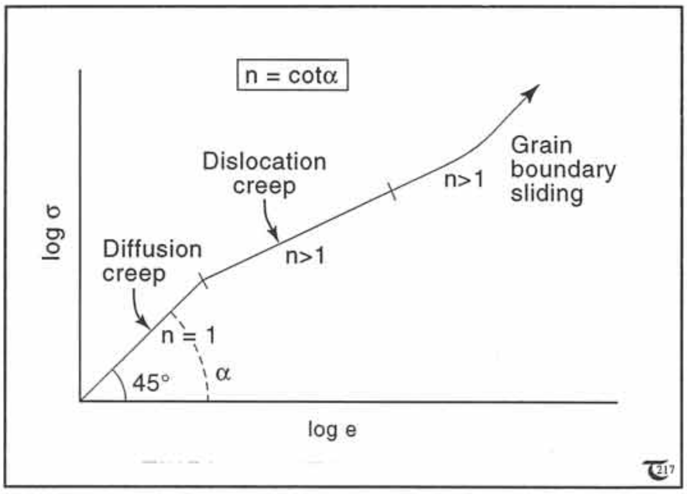


Figure 7-23: Regimes of diffusion creep, dislocation creep, and superplasticity (grain boundary sliding), as they appear in a log stress-log strain-rate plot. The n-value, characteristically constant within each regime of flow, is determined by the slope of the flow curve for steady-state creep. The logarithmic scales for stress and strain both need to be scaled equally for determination of the n-value.

data of hot press tests can be visualized by plotting pairs of steady-state stress and strain-rate to obtain *log stress-log strain-rate* curves, as plotted for Yule Marble in Figure 7-22. The positive slope for the flow curves occurs only because the strain-rate scale increases toward the right-hand side. The slope of the flow curves commonly appears positive when the strain-rate is plotted to increase in the positive X-direction, as illustrated in Figure 7-23.

The shape and slope of steady-state curves in *log stress-log strain-rate* plots can be used to distinguish whether creep is by diffusion, dislocation movement, or superplasticity. Figure 7-23 shows an idealized graph, indicating that slope unity corresponds to diffusion creep, a slope larger than unity corresponds to dislocation creep, and unstable slopes are characteristic for superplasticity at higher strain-rates. *Superplasticity* is thought to be largely due to a combination of grain-boundary sliding, assisted by internal distortion of crystal grains by dislocation creep, and generally requires $T_H > 0.5$.

☐**Exercise 7-8: Prepare a *log stress/log strain-rate* plot on isometric log scales, using part of the curves in Figures 7-21a to c, where these approach steady-state flow. Determine the likely deformation mechanism that allowed the flow.**

7-7 Deformation maps

The shape of flow curves and the corresponding microstructural mechanisms have been determined for many common minerals and some monomict and polymict crystalline rocks. A *deformation map* or *Ashby plot* provides a comprehensive way to visualize the flow mechanisms of either a particular mineral or rock. Deformation maps show the relationship among the differential stress, strain-rate, and either the grain size or the temperature for a given rock or mineral.

Deformation maps, also, visualize which microstructural deformation mechanism dominates under which physical conditions (Figs. 7-24a & b). The important effect of grain-size reduction as a softening mechanism has been separately graphed in Figure 7-25.

Deformation maps may provide useful supplementary information to field geologists. They can be used to estimate, to some extent, the range of strain-rates at which the observed microstructure and the corresponding field structure developed. If the strain-rate is already inferred from field observations and the microstructure reveals the dominating deformation mechanism, then it is possible to constrain the range of tectonic stresses at which the deformation took place. Nonetheless, deformation maps should be applied with care and some reserve. As more practical examples appear in the literature, the better will their significance be understood.

☐**Exercise 7-9: It is well known that grain-size reduction takes place during dynamic recrystallization. Assume now that the grain size of quartz reduces from 1 mm to 0.01 mm. Use the plot of Figure 7-25 to infer:**
(a) What happens to the stress if the deformation continues at a constant strain-rate of 10^{-14} s^{-1} ? (b) What happens to the strain-rate if the deformation were to continue under a constant stress of 1 MPa?

7-8 Ductile versus plastic creep

The term ductile here refers to deformation in the regime of crystalline creep. Propositions to use the term *plastic* instead of *ductile* have gained support by some geoscientists. However, the term plastic is commonly used in continuum mechanics to describe deformation, where the stress input is similar for all displacement rates and no reference is made to the physical texture of the material or

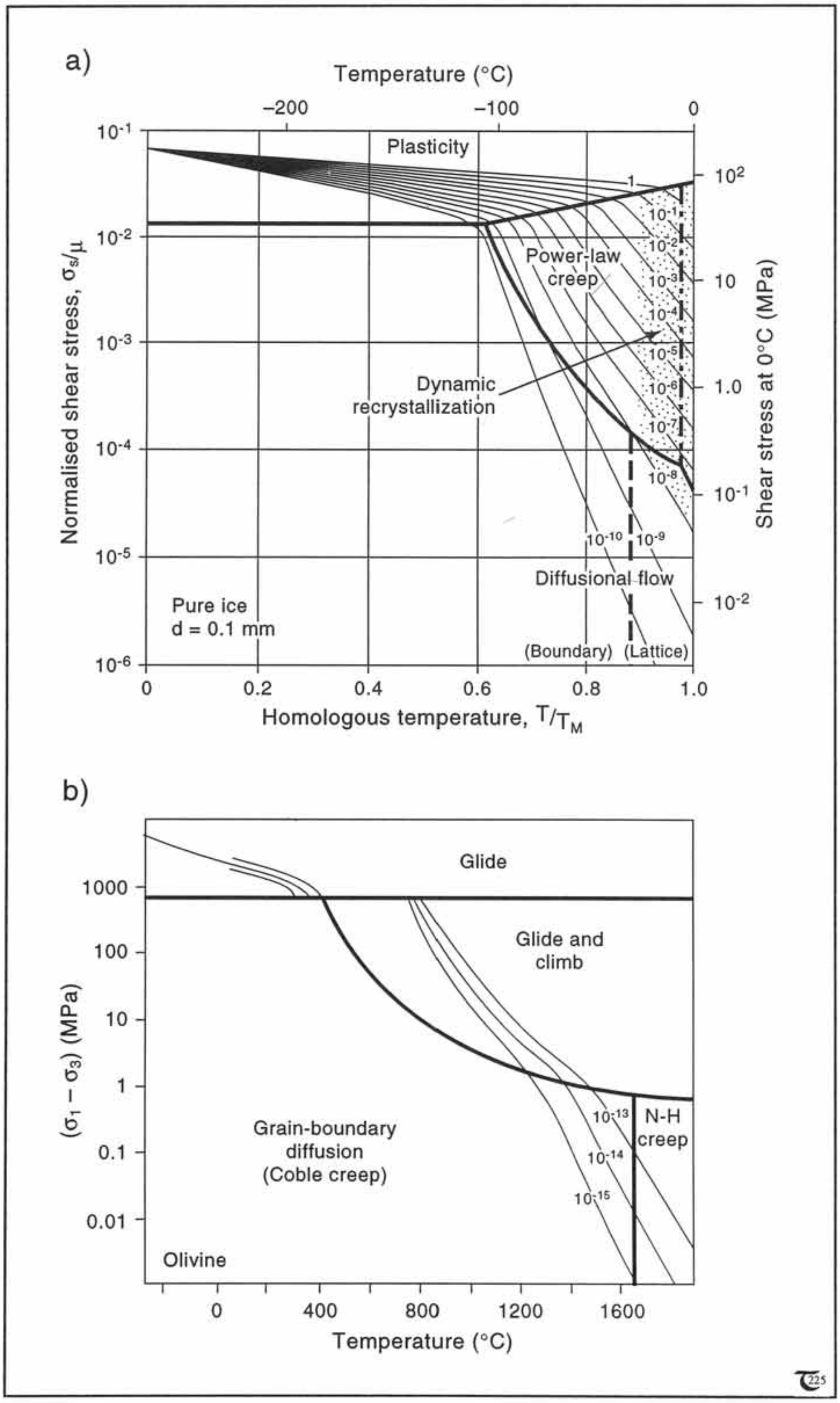

a)
Temperature (°C)
−200
−100
0
Plasticity
Power-law creep
Dynamic recrystallization
Diffusional flow
(Boundary) (Lattice)
Pure ice
d = 0.1 mm
Normalised shear stress, σs/μ
Shear stress at 0°C (MPa)
Homologous temperature, T/TM
b)
Glide
Glide and climb
Grain-boundary diffusion (Coble creep)
N-H creep
Olivine
(σ1 − σ3) (MPa)
Temperature (°C)

Figure 7-24: *a) & b) Examples of deformation on Ashby maps for pure ice (a) and olivine (b). Each point in the deformation map is determined by a temperature, stress, and strain-rate. What follows is the creep mechanism prevailing under those conditions. Note that each map is valid only for a particular grain size, as the deformation fields shift with varying grain sizes.*

to the process supporting such a particular mode of deformation. Mineral grains may deform by *crystal plasticity* in some mechanisms of creep, but the term *ductile* is clearly preferrable to qualify the global flow. This is because the rate of the global flow is generally stress-dependent and, therefore, cannot be termed plastic. (Perfect plasticity requires an n-value of ∞, so that the stress remains constant for whatever strain-rate occurs.) Additionally, the stress state and brittle failure of rocks in the shallow crust can best be described in terms of the theory for *plastic failure* and *frictional plasticity*. No artificial separation or confusion should be encouraged between *frictional plasticity* and *crystal plasticity*, which would occur if the term plastic were to be reserved for one of these processes only.

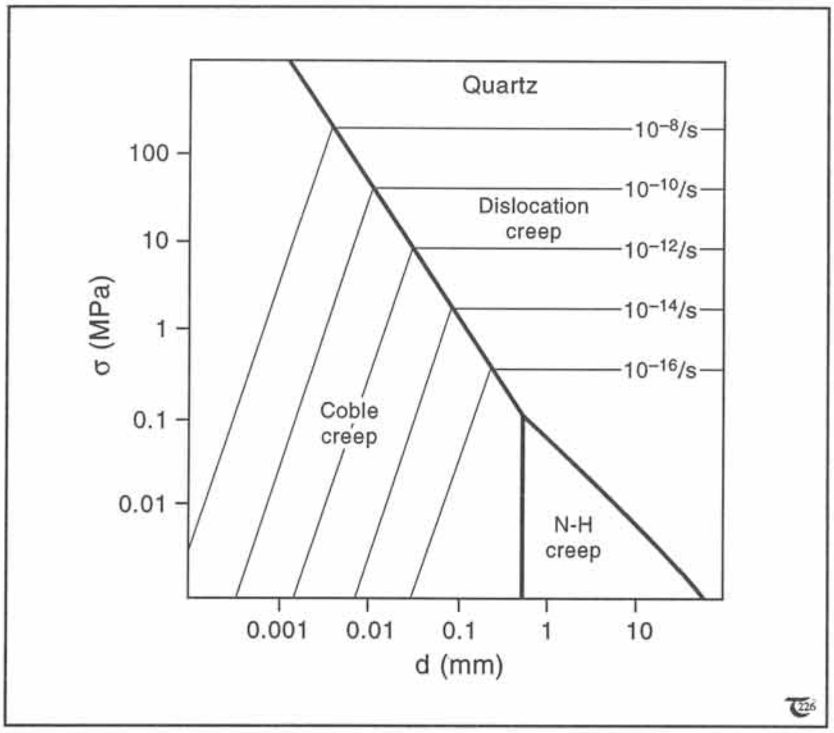


Figure 7-25: *Plot of deformation mechanisms for quartz, showing the importance of grain size for the mechanism of creep. See exercise 7-8.*

References

A. Books

Deformation-Mechanism Maps (1982, Pergamon Press, Oxford, 166 pp), by H.J. Frost and M.F. Ashby. A landmark text with many examples of experimental creep data, generalized in deformation maps for a variety of rocks and minerals.

Creep of Crystals (1985, Cambridge University Press, Cambridge, 260 pp), by J.P. Poirier. This is a very instructive introduction to the detailed processes of crystalline creep.

B. Articles

The following articles provide useful complementary reading to the topics discussed in this chapter:

Kerrich, R. and Allison, I. (1978, *Geoscience Canada*, volume 5, pages 109 to 118). Flow mechanisms in rocks.

Means, W.D. (1989, *Journal of Structural Geology*, volume 11, pages 163 to 174). Synkinematic microscopy of transparent polycrystals.

Schedl, A. and Van der Pluijm, B.A. (1988, *Journal of Geological Education*, volume 36, pages 111 to 121). A review of deformation microstructures.

Schmid, S.M. (1986, Chapter 1.8 in: Hsu, K., editor, *Mountain Building Processes*). Microfabric studies as indicators of deformation mechanisms and flow laws operative in mountain building, pages 95-110.

Groshong, R.H. (1988, *Geological Society of America Bulletin*, volume 100, pages 1329 to 1360). Low-temperature deformation mechanisms and their interpretation.

C. Videotapes

Development of S-C Structures and Grain-boundary Sliding Superplasticity (1988, University of Michigan, Deptartment of Geological Sciences, USA, 15 mins.), by Ben van der Pluijm.

Deformation of Mineral Analogs under the Microscope (1986, University of Utrecht, Geology Department, Netherlands, 20 mins.), by Janos Urai.

Dynamic Processes in Shear of Ice as a Rock Analogue (1986, University of Melbourne, Centre for the Study of Higher Education, Parkville, Victoria 3052, Australia, 30 mins.), by Chris Wilson.