

Some Current Aspects of Experimental Rock Deformation [and Discussion]

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Phil. Trans. R. Soc. Lond. A 1976 **283**, 163-172

doi: 10.1098/rsta.1976.0076

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Some current aspects of experimental rock deformation

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This review considers first some general aspects of the relevance of laboratory rock deformation experiments to geological conditions and then, specifically, recent work on olivine and quartz. Improvements in temperature capability and reliability of apparatus now permit the exploration of regimes of flow characteristic of higher temperatures and lower stresses and strain rates than those used hitherto. These experimental advances, together with the availability now of transmission electron microscopy, open the way to a greater understanding of the major deformation mechanisms, an understanding, or at least recognition, that must form a vital link in validly extrapolating experimentally determined flow laws to natural conditions. Recent work on olivine crystals has considerably improved our understanding of its deformation mechanisms and likely geological behaviour, but in the case of quartz there are many aspects of behaviour, especially involving the role of water, that remain obscure.

INTRODUCTION

In giving this brief review of the 'state of the art' in a field that has become quite a large one, it is necessary to limit severely the scope attempted. In doing so, the choice has been made of some very general aspects of experimental rock deformation, followed by comments on the status of research on the deformation of olivine and quartz and their rocks. One aim underlying this approach is to attempt to set down and illustrate the rationale of the experimental studies, especially in relation to their geological relevance.

In general, experimental work on the deformation of minerals and rocks aims at elucidating their mechanical properties, including the processes of deformation and the related structural changes. That is, the aim is to characterize the minerals and rocks *as materials* in relation to their mechanical behaviour. This work is essentially distinct from but complementary to experimental and computer simulation studies with models which attempt to solve boundary value problems of geology. In the latter studies, a model material is chosen or assumed which is supposed to scale the flow laws of the rocks concerned at the same time as the dimensions, displacement rates and so forth are being scaled. But the question remains of how to discover what are the flow properties of rocks that apply on the geological time-scale, under the geological conditions of temperature, pressure and tectonic stresses.

THE EXTRAPOLATION PROBLEM

The experimental approach to determining the flow properties of rocks under geological conditions involves extrapolation from laboratory conditions, notably in time-scale but correspondingly in other parameters too. But the validity of extrapolation inevitably involves questions of the mechanism of deformation. For there is no point in attempting an extrapolation unless we know or can demonstrate in some way that the flow process involved in the

experiments from which we are extrapolating is the same one as is involved on the geological time-scale; otherwise there are no grounds for expecting the same flow law to hold throughout.

This point can be illustrated schematically with a diagram (figure 1) depicting a hypothetical, but not unlikely, situation in which at a given geologically interesting temperature T_1 a flow process II that occurs more readily than a process I on the geological time-scale is inaccessible to experiment on the laboratory time-scale because at the strain rates measurable in the laboratory the flow stress for process I is lower than for process II. However,

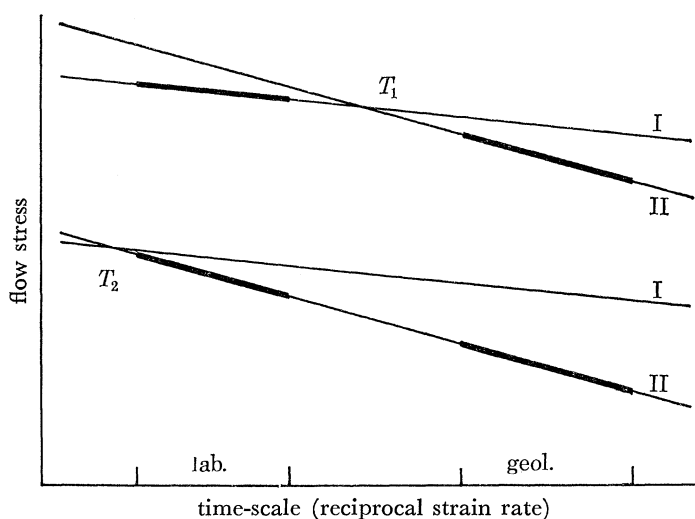


FIGURE 1. Hypothetical dependence of flow stress on strain rate for two deformation mechanisms I and II at temperatures T_1 and T_2 ($T_2 > T_1$). The emphasized portions of the lines indicate the active mechanism on the respective laboratory or geological time-scale.

the experience with many materials is that raising the temperature has a greater influence on the more strain-rate sensitive process, so that the cross-over of the flow stress curves will occur at shorter times at higher temperatures. In this case, as shown in figure 1, raising the temperature suitably can make the process II accessible to study on the laboratory time-scale. It is now possible to determine the temperature and strain-rate dependences of the flow stress over a limited field and so establish relationships through which extrapolation can then be made to the lower strain-rates and lower temperatures of geological interest under the assurance that the same flow process is operative over the range of extrapolation.

The extrapolation procedure just outlined is, of course, rather idealistic and many difficulties arise in practice. Thus, melting or other phase changes limit the extent to which higher temperatures can be exploited, there tend to be more than just two flow processes to take into account, and there are practical difficulties in establishing that one and the same process has been effective over the whole range of conditions covered in the extrapolation. The variety of deformation mechanisms and some of the practical matters will now be touched upon in turn.

THE MULTIPLICITY OF DEFORMATION MECHANISMS

Mechanisms of deformation that are possibly of importance in rocks can be conveniently grouped into three categories:

(a) Cataclastic

‘Cataclastic flow’ is a term largely peculiar to the literature of rock deformation: it refers to permanent straining achieved by the combination of a distributed microfracturing, whereby the material is effectively fragmented, and relative movement on the fractures, somewhat analogous to the flow of sand. Two important properties of cataclastic flow are a strong dependence on confining pressure and a tendency for volume changes to occur, while the temperature dependence can be expected to be relatively small. Purely cataclastic flow may be rare but a component of cataclastic deformation, at least to the extent of limited microcracking, may be not uncommon at relatively small depths and low temperatures, with important implications for permeability and phenomena involving pore fluids.

(b) Crystal-plastic

Crystal plasticity is used here to cover any change of shape achieved predominantly by the glide motion of dislocations, as expressed in crystallographic slip or twinning. No substantial volume changes are involved and so only relatively small effects of confining pressure are to be expected. The resistance to deformation (flow stress) resides ultimately in the difficulty of moving the dislocations but there are many different factors that may predominate under different circumstances in determining this resistance. The factors include the intrinsic difficulty of moving a dislocation through the crystal structure (measured by the Peierls stress), the difficulty of cutting through or circumventing precipitates or cross-cutting dislocations, and the counteracting of the ‘back stress’ arising from the stress fields of other dislocations in the neighbourhood. Temperature change will tend to influence these various factors in different ways but in general it is useful to distinguish between regimes where the factor determining the flow behaviour is relatively insensitive to temperature (e.g. where the flow stress is determined primarily by the internal stress field from other dislocations) and regimes where the flow is strongly dependent on temperature (e.g. where the strain rate is determined by the rate at which obstacles are surmounted by the dislocations climbing into adjacent slip planes or where the Peierls stress plays an important part). The way in which the strain rate depends on the level of the stress will also tend to vary with the nature of the resistance to dislocation motion. Thus, there are a variety of flow laws that may be expected to apply under different circumstances even when the strain is always produced by crystallographic slip.

(c) Diffusional

Diffusional flow is taken to include any process whereby the diffusion of individual atoms or groups of atoms occurs in such a way as to produce a net change in the shape of the body. The form of the flow law will depend on the details of the diffusion path or mechanism, which may involve diffusion through the grains (Nabarro-Herring) or along grain-boundaries (Coble) or subgrain-boundaries, the pure climb motion of dislocations (Nabarro) or diffusion through an interstitial fluid phase. In general, all these processes can be expected to be highly dependent on temperature but relatively insensitive to pressure, except perhaps for pore pressure in the last case. For further details of both diffusional and crystal-plastic processes, see Weertman

(1970, 1975), Weertman & Weertman (1970, 1975), Kirby & Raleigh (1973), and Stocker & Ashby (1973).

Because of the above multiplicity of possible deformation mechanisms and because different ones are likely to be found under different external conditions, it is useful to have ways of representing this variation with conditions. The '*deformation mechanism map*' of Ashby (1972) is one such representation that is particularly useful in visualizing the relationships between mechanisms. It is based on the presumption that, where there are competing mechanisms under given physical conditions, one mechanism will in general predominate over the total activity of all others within a certain field of conditions. In this way, fields of conditions for the predominance of various mechanisms are defined, as illustrated in figure 2. The strain rates for steady-state flow under given stress and temperature can also be represented on the diagram by strain rate contours, as shown. It should be noted that such a diagram refers to a specific grain size and that the boundaries of the fields (especially of Coble and Nabarro-Herring creep) may move significantly with change in grain size.

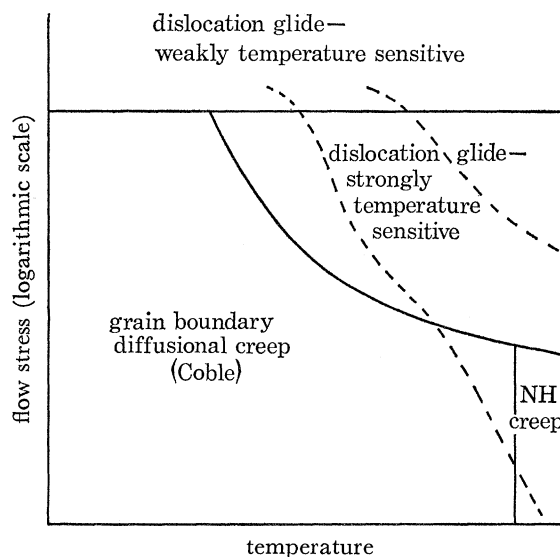


FIGURE 2. A schematic deformation mechanism map, based on that predicted by Stocker & Ashby (1973) for olivine-rich rocks. The areas delineated represent stress-temperature regimes in which the particular deformation mechanisms designated predominate. The dashed lines represent possible contours of constant strain rate for steady state creep.

The use of stress as a primary parameter determining the mechanism of deformation is especially worth remarking upon. It reflects a wide experience with metals and ceramic oxides and if, as seem likely, it holds also for rocks, important implications follow both for conducting appropriate experiments and for interpreting field observations. Thus, in doing experiments to obtain flow data for extrapolation to geological conditions, it is important to choose stresses and temperatures that are in the same mechanism field as the geological conditions. In practice, this means that there may well be an upper limit of fairly modest value to the stresses that can be used and so in order to achieve strain rates that are appreciable on the laboratory time-scale, temperatures considerably higher than those in nature may have to be used. There are, of course, upper limits also on the temperatures that can be used, set either by other boundaries of the mechanism fields or by phase stability boundaries, but it is important to push the

experiments to as high temperatures as possible above those occurring in nature in order to be as sure as possible of establishing mechanism fields relevant to the natural situations. A good deal of the experimental work done so far has probably been at too low temperatures and consequently at too high stresses to represent fully the geological flow processes and caution is needed in extrapolating flow data from such experiments.

PRACTICAL ASPECTS

The continuing development in rock deformation apparatus is contributing substantially to meeting the need for higher temperatures just emphasized. High temperatures at high pressures are most readily produced in solid-medium apparatus of the Griggs type (Griggs 1967), which has largely dominated the experimental scene at high temperatures in recent years. However, such apparatus is at a disadvantage for stress-strain and creep studies at low stresses because of problems with friction and the strength of the medium itself, although some improvements have been made with the use of weaker pressure media such as salt or glass and the use of an internal load cell (Blacic & Hagman 1974). Therefore, for quantitative work on flow laws, it is highly desirable, even possibly essential, to use gas medium apparatus with internal furnace and internal load cell gauge, such as the apparatus of Heard (Heard & Carter 1968) and of the author (Paterson 1970). The use of an internal load cell is normally required for work at low stresses because of the vagaries of friction. Also for high temperature work it is essential to place the furnace internally; in the author's experience (Paterson 1975), this enables temperatures of at least 1300 °C to be reached (and it is believed that the present design is capable of appreciably higher temperatures), with a confining pressure of at least 300–500 MPa (3–5 kbar). Some confining pressure is normally important for the polycrystalline rocks, but it is doubtful whether pressures higher than 300–500 MPa are required in order to achieve the types of relatively low-stress deformation process that are of geological interest. However, a larger pressure range could be useful for establishing accurately the pressure dependence for extrapolating results to deep crust and upper mantle conditions, or for working with phases that are only stable at higher pressures.

Another practical question is that of how to recognize what deformation mechanism is predominating or at least to establish that the same process is occurring in the specimen as has occurred in nature. Here we are largely limited to microstructural observation because this is the nature of the record remaining of the geological behaviour; nevertheless it is a large subject that cannot be reviewed adequately here. The optical microscope has played and can be expected to continue to play a very important part, as will X-ray fabric determinations, but the application of the transmission electron microscope to minerals has greatly extended the horizon on microstructural observation. Now the dislocations themselves, the basic entities involved in many of the conceivable deformation mechanisms, can be observed directly and their Burgers vectors and other characteristics determined (for example, Ardell, Christie & McCormick 1974). Thus, the clues to the operation of identical processes in nature and in the laboratory must lie in the observation of similar microstructures produced. But in applying this principle, it must be borne in mind that new techniques may have to be developed to reveal relevant aspects of the microstructure that are not normally studied. For example, the detection of deformation by diffusional transfer of material may call for novel techniques such as micro-analysis for minor elements on the electron microscope scale.

STUDIES PERTAINING TO OLIVINE-RICH ROCKS

Recent studies have centred on olivine single crystal behaviour, partly on the grounds that the essential behaviour of olivine-rich rocks is already revealed in that of olivine single crystals; in this case high temperatures can be exploited that would normally be above the solidus of the polymineralic rock (Kohlstedt & Goetze 1974). Deriving flow data for the rocks from single crystal measurements, however, carries the implication that, since the von Mises criterion for ductility is not satisfied by the known slip systems of olivine (Paterson 1969), the inadequacy in slip systems can be compensated for by heterogeneity of deformation on the scale of the grains or by minor activity of other deformation mechanisms without modifying the flow parameters relating to the individual grains.

The recent single crystal work begins with the electron microscope study on deformed specimens by Phakey, Dollinger & Christie (1972; for a review of this and earlier work, see Paterson 1974). However, more comprehensive and pertinent flow stress data come from the experiments of Kohlstedt & Goetze (1974) and Kohlstedt, Goetze & Durham (1975) which have been done at atmospheric pressure with controlled oxygen fugacity at higher temperatures, in the range 1400–1650 °C. The direction taken by this work illustrates well the argument developed above that laboratory studies of geological relevance need to be done at temperatures higher than those of geological interest. In this case, geologically interesting temperatures are already high, perhaps 1200 °C for upper mantle flow, but even at this temperature high stresses are required for deformation in the laboratory and the mechanical behaviour still involves high work hardening (for example, an unpublished experiment by the author shows that in a synthetic pure forsterite crystal, compressed at 45° to the *a* and *c* axes at 1200 °C and 300 MPa confining pressure, the flow stress rises from about 140 to 300 MPa in the first 7 % of strain at a constant strain rate of 10^{-5} s^{-1}). On the other hand, Kohlstedt & Goetze find steady state creep with strain rates of around 10^{-6} s^{-1} at stresses of 10 MPa or less at their highest temperatures. These stresses are much lower than those used in previous olivine studies. It is notable that the dislocation structures are also markedly different from those produced under high stresses and Kohlstedt & Goetze argue that they are similar to naturally occurring dislocation structures which may then be related to flow under similarly low stresses in nature, although at lower temperatures at which strain rates will be correspondingly lower. It is also notable that in this regime the strain rate in creep is found to be less strongly stress-dependent than at higher stresses, so it would appear that one has crossed into a different field in the deformation mechanism map, a field that is probably geologically more relevant, at least for upper mantle flow.

STUDIES PERTAINING TO QUARTZ-RICH ROCKS

As with olivine, recent studies have largely focused on single crystal behaviour and discussion will be limited to this, but in the case of quartz the work has been concerned with a number of fundamental qualitative questions arising out of the discovery of water weakening (Griggs & Blacic 1964, 1965; Griggs 1967). Since the time of this discovery, which was made on natural crystals that were exposed to water from the dehydration of talc at 800 °C and above at 1500 MPa confining pressure, all single crystal studies involving substantial amounts of plastic deformation have been done on synthetic crystals except for some experiments at

1020 °C, 800 MPa by Heard & Carter (1968). Thus there is still very little information on the stress–strain properties of natural quartz crystals except that they have yield stresses of the order of the ‘theoretical strength’ in the temperature range 300–700 °C (Griggs & Blacic 1964) but much less at 1020 °C (Heard & Carter 1968). However, the experiments on natural crystals have given much information on a multiplicity of slip systems and associated microstructures (Carter, Christie & Griggs 1964; Christie, Griggs & Carter 1964; Christie & Green, 1964; McLaren, Retchford, Griggs & Christie 1967; Heard & Carter 1968; Christie & Ardell 1974).

The studies on synthetic quartz have been on material that contains substantial amounts of OH incorporated during growth, as indicated by infrared absorption in the neighbourhood of 3 μm wavelength. Following on from the measurements done with solid-medium apparatus (for the bulk of these, see Blacic 1971, 1975), there is by now a considerable body of observations made in gas-medium apparatus or at atmospheric pressure (Heard & Carter 1968; Baëta & Ashbee 1970, 1973; Hobbs, McLaren & Paterson 1972; Balderman 1974; Morrison-Smith, Paterson & Hobbs 1975). Two important aspects of the stress–strain behaviour and associated dislocation structures emerge from this work.

(a) The water-weakening effect includes, in addition to a reduction in yield stress, a very marked reduction in the work-hardening rate with small increases in temperature in the neighbourhood of the hydrolytic weakening temperature defined by Griggs & Blacic from relaxation tests. This change is accompanied by a marked change in the character of the dislocation structures developed, from loops showing strong crystallographic control or dense tangles of dislocations to lower density configurations of curved and intertwined dislocations associated with dipoles and ‘debris’ loops.

(b) In many cases, there is a pronounced yield point drop analogous to that observed in lithium fluoride, germanium and similar materials. The yielding is also accompanied by an extremely marked increase in dislocation density.

The yielding behaviour and the indications of strong crystallographic control on dislocations at the lower temperatures (presumably reflecting a high Peierls stress) strengthen the feeling that, through a similar nature of bonding, quartz will have much in common in mechanical behaviour with germanium and silicon. It was therefore suggested by Hobbs, McLaren & Paterson (1972) that the microdynamical theory developed for these materials by Haasen (1964*a, b*; see also Alexander & Haasen 1968) could be applied to the yielding behaviour of quartz and that it could probably also be developed to describe the whole stress–strain curve. Such a development has been attempted by Griggs (1974), who has proposed a microdynamical model for the mechanical behaviour of hydrolytically weakened quartz which includes a specific role for the water and incorporates recovery effects. This model reproduces to quite a remarkable degree much of the complexity of the existing stress–strain data. However, for further progress in testing this or alternative models, more precise stress–strain and creep data are needed and, more importantly, separate observations relating directly to specific aspects of the models, in particular the laws governing the velocity and multiplication of dislocations and the strain-hardening and recovery processes.

The experiments on synthetic quartz have, however, shown up a number of difficulties in working with this material, some of which are well illustrated in the following observations in a recent study (Morrison-Smith, Paterson & Hobbs 1975):

(a) The OH content was markedly heterogeneous, being banded parallel to the growth faces and giving rise to wide scatter in results and marked heterogeneity in deformation of individual

specimens, as has also been noted by others (Baëta & Ashbee 1970; Blacic 1971, 1975); and it is very difficult to homogenize synthetic crystals by annealing them.

(b) Submicroscopic entities, thought to be precipitates, were present in the crystal used in this study and they had a very pronounced influence on dislocation multiplication.

(c) Above about 800 °C, water was precipitated, as indicated by the appearance of small bubbles visible in the electron microscope (cf. also Baëta & Ashbee 1973, and Balderman 1974).

It would now appear desirable to return to the experimental study of natural quartz crystals, especially if water could be introduced in a homogeneous way. However, attempts by the author to introduce water into 5 mm diameter jacketed natural quartz specimens in a gas apparatus at 300 MPa confining pressure and 1000 °C by putting some water inside the sealed jacket and heating for 2 h have been unsuccessful in producing marked weakening, even when the water was introduced by dehydrating a talc sleeve as in the experiments of Griggs & Blacic (it may be noted that a different crystal from theirs was used for specimens). In order to resolve this dilemma and to make further progress in understanding the role of water in quartz deformation in general, it is desirable to have much more information on the solubility and diffusion of water or its components in quartz, covering the influence of pressure, temperature and concentration, including particularly that of other impurities (a private communication from A. H. Heuer, A. R. Cooper & J. Sang indicates that other impurities, especially aluminium, may profoundly affect the rate of diffusion). It is also highly desirable to gain more information on the chemical state of the 'water' within the quartz in the various experimental situations, as revealed, for example, by infrared absorption. In conclusion, therefore, it is seen that there are many aspects of the deformation of quartz and of the role of water in it that are at present poorly understood and in need of detailed further work, with natural crystals as well as synthetic.

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Discussion

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I agree fully with Dr Paterson's suggestion, that there is still much to learn about the mechanical properties of quartz. Synthetic quartz is a very complex material and is far from being ideal for use in experimental deformation. The concentrations of hydroxyl ions in such crystals show a complex distribution, not only as a result of banding in the Z-growth region, but also because the rate of hydroxyl ion incorporation varies for different growth sectors. Observations of the temperature of initial formation of bubbles due to precipitated water, have shown that in all cases investigated the prism growth sector has turned milky at a lower temperature than any other. These observations also indicate that in the Z-growth sector, the temperature of initial formation of bubbles is inversely proportional to the hydroxyl concentration measured before heat treatment. Those growth bands with the highest hydroxyl concentration turning milky at the lowest temperature. If this applies to other growth sectors, the maximum hydroxyl concentration is found in the prism.

In the synthetic crystals used in this work, the growth rate of the prism is slow compared to that of the Z-growth region, so the established axiom, that the hydroxyl concentration is proportional to the growth rate, cannot be applied in a general manner. Prism growth is more important in natural than in synthetic quartz crystals, and it might therefore be expected that any natural quartz crystal, which had grown rapidly, would support a very high hydroxyl concentration. Infrared studies of quartz from low temperature tectonic veins have shown that this is indeed the case. Measured concentrations varying from 5000 to 12000 H/10⁶ Si. Such quartz might be expected to be mechanically weak.

Samples of synthetic and natural (Brazilian) quartz crystals were heated to temperatures in the range 200–1000 °C for periods of between 5 h and 7 days. Infrared studies of the samples indicate that the concentration is temperature dependent, the crystals dehydrating at temperatures greater than 300 °C following an Arrhenius type relationship. Equilibrium appears to be established after about 12 h for a given temperature, although different crystals follow different equilibrium paths. I suggest that this effect may be important in deformation experiments, especially if the concentration dependence of the weakening temperature is being investigated. One of the synthetic crystals used in this study, with an initial concentration of 2500 H/10⁶ Si,

was found from deformation experiments to have a weakening temperature of about 475 °C. However, the sample used to determine this was heated under test conditions for 24 h before loading to ensure that the hydroxyl concentration had equilibrated. The equilibrium concentration for this crystal at this temperature is only 832 H/10⁶ Si, less than a third of the concentration measured before the test. Dr Paterson suggested that at higher pressures, this dehydration might be retarded and that its effects may then be ignored. I have not as yet investigated the influence of pressure on the dehydration curve, but as bubbles of water are precipitated in samples during deformation tests, I expect the effect to be slight. This is currently being investigated.