Dislocation and diffusion creep of synthetic anorthite aggregates

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Abstract. Synthetic fine-grained anorthite aggregates were deformed at 300 MPa confining pressure in a Paterson-type gas deformation apparatus. Creep tests were performed at temperatures ranging from 1140 to 1480 K, stresses from 30 to 600 MPa, and strain rates between 2×10^{-6} and 1×10^{-3} s⁻¹. We prepared samples with water total contents of 0.004 wt % (dry) and 0.07 wt % (wet), respectively. The wet (dry) material contained <0.7 (0.2) vol % glass, associated with fluid inclusions or contained at triple junctions. The arithmetic mean grain size of the specimens varied between $2.7 \pm 0.1 \,\mu\text{m}$ for the dry material and $3.4 \pm 0.2 \,\mu\text{m}$ for wet samples. Two different creep regimes were identified for dry and wet anorthite aggregates. The data could be fitted to a power law. At stresses >120 MPa we found a stress exponent of n = 3 irrespective of the water content, indicating dislocation creep. However, the activation energy of wet samples is 356 ± 9 kJ mol⁻¹. substantially lower than for dry specimens with 648 ± 20 kJ mol⁻¹. The preexponential factor is log A = 2.6 (12.7) MPa⁻ⁿ s⁻¹ for wet (dry) samples. Microstructural observations suggest that grain boundary migration recrystallization is important in accommodating dislocation creep. In the lowstress regime we observed a stress exponent of n = 1, suggesting diffusion creep. The activation energies for dry and wet samples are 467 ± 16 and 170 ± 6 kJ mol⁻¹, respectively. Log A is 12.1 MPa⁻ⁿ μ m^m s⁻¹ for the dry material and 1.7 MPa⁻ⁿ μ m^m s⁻¹ for wet anorthite. The data show that the strengths of anorthite aggregates decrease with increasing water content in both the dislocation and diffusion creep regimes. A comparison of the creep data of synthetic plagioclase from this study with published data for feldspar, olivine, and quartz indicates a linear relationship between activation energy and log A similar to the suggested compensation law for diffusion in silicates.

1. Introduction

Geological field studies and laboratory investigations of the mechanical properties of rocks suggest a rheological stratification of the continental lithosphere [e.g., Brace and Kohlstedt, 1980; Kohlstedt et al., 1995] with a weak lower crustal layer sandwiched between stronger middle crust and upper mantle [e.g., Ranalli, 1987]. Independently, Meissner and Strehlau [1982] and Chen and Molnar [1983] suggested that reduced seismic activity of the lower crust indicates low ductile flow strength relative to the upper crust. However, mechanical data and flow law parameters on the rheology of rocks typical for the lower crust, i.e., gabbros, metabasites, etc., are still scarce [Caristan, 1982; Shelton and Tullis, 1981; Kirby, 1983; Carter and Tsenn, 1987; Mackwell et al., 1998]. The amount of stress transfer and partitioning of strain between crustal layers and the mantle has given rise to much controversy [e.g., Kirby, 1985; Ord and Hobbs, 1989].

Feldspars, pyroxenes, and amphiboles are the most abundant minerals constituting the deep continental crust. For monomineralic rocks composed of feldspar or pyroxene some recent studies of synthetic aggregates exist that yield constitutive equations for diffusion creep [Lavie, 1998; Dimanov et al., 1998, 1999]. Bystricky [1998], Mackwell et al. [1998], and Xiao [1999] investigated mixtures of clinopyroxene or quartz with feldspar at various volume fractions. Detailed studies of the microstructure

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Paper number 2000JB900223. 0148-0227/00/2000JB900223\$09.00 and inferred deformation mechanisms exist for experimentally and naturally deformed feldspathic and quartzo-feldspathic rocks and are discussed in a seminal review by *Tullis* [1990].

The high-temperature plastic strength of rocks can be reduced considerably by the presence of water-related species [Kohlstedt et al., 1995]. Tullis and Yund [1980, 1991] and Tullis et al. [1996] showed in their experiments that the presence of water reduces the strength of feldspar and may cause a change of the dominant deformation mechanism. Dimanov et al. [1999] demonstrated that the strength and activation energy for grain boundary diffusion-controlled creep of pure fine-grained anorthite aggregates is reduced at atmospheric pressure when traces of water (0.05-0.1 wt %) are present. Likewise, the stress at the transition from diffusion to dislocation creep of feldspar is lower with trace amounts of water present.

In this study we investigate dislocation and diffusion creep of anorthite aggregates at elevated confining pressures. We present constitutive flow laws for synthetic anorthite aggregates that are dry or contain trace amounts of water.

2. Starting Material and Experimental Methods

2.1. Specimen Fabrication

Anorthite aggregates were fabricated from crushed CaAl₂Si₂O₈ glass (Schott Glaswerke). The chemical composition of the glass from fluorescent X-ray analysis is An_{98 9}Or_{0.2}Ab_{0.9} (normalized to 8 O) with major impurities of TiO₂ (0.1 wt %), MgO (0.07 wt %), Fe₂O₃ (0.01 wt %), and H₂O (\approx 0.05 wt %).

The particle size of the glass powder was <60 μ m. The powder was predried at 1073 K in air for 60 hours and kept at 393 K for

several days before cold pressing into steel jackets with 0.5 mm wall thickness. A vacuum dye was used to reduce pressure to 1.5 kPa. To cold press the samples, an axial stress of 300 MPa was applied. The diameter of the jackets was 15 mm and the length was 30 mm. Using a gas pycnometer (Micromeritics Accy Pyc 1330), the porosity of the green bodies was determined to be ≈30%. To fabricate "dry" specimens, we heat treated the coldpressed pellets at 1073 K to 1173 K for more than 120 hours prior to hot pressing. "Wet" samples were hot isostatically pressed without the initial heat treatment. A Paterson-type gas apparatus was used to hot press and anneal both wet and dry samples in three steps at a pressure of 300 MPa. First, samples were further densified above the glass transition temperature at 1173 K for 1 hour. Further annealing was performed at 1323 K for 2 hours, i.e., slightly above the nucleation temperature of 1297 K for anorthite [Dresen et al., 1996]. Finally, the annealing temperature was increased to 1473 K for 2 hours to promote grain growth. Heating and cooling rates were 20 and 10 K min⁻¹, respectively. After samples were retrieved from the vessel, the iron jacket was dissolved in acid (50/50 vol % HCl/HNO₃). For mechanical testing, the samples were precision ground to obtain cylindrical specimens of 10 mm diameter and 20 mm length. In this procedure the initial sample surface possibly contaminated by the acid was removed, and the sample diameter was reduced by 4 mm. The porosity of the samples is less than $\approx 1\%$, as determined by Archimedes' method in water.

2.2. Water Content

A Fourier-transformed infrared spectrometer (FTIR, Bruker IFS-66v) equipped with a microscope was used to measure the water content of samples before and after deformation. The spectra were recorded at 293 K from >128 scans between wavenumbers 2500 to 4000 cm⁻¹ (resolution of 2 cm⁻¹). We performed at least five separate measurements per sample. Since the spot size of the microscope is roughly 20 times larger than the average anorthite grain size, our measurements refer to the total water content of the samples. To quantitatively estimate the water content of the aggregates from 150-µm-thick samples, we used Beer-Lambert's law and a molar extinction coefficient of 32 $L_{(H2O)}$ mol⁻¹ cm⁻¹ [Beran, 1987]. This technique probably yields an upper limit for the water content [Bell et al., 1995].

Absorbance spectra for typical wet and dry samples are given in Figure 1. The spectra show a broadband absorbance with a maximum near 3600 cm⁻¹, characteristic of molecular water or hydroxyl [Aines and Rossman, 1984]. Specimens referred to as wet contain about 11500 \pm 2900 ppm H/Si (0.07 \pm 0.02 wt % H₂O) on average and 640 \pm 260 ppm H/Si (0.004 \pm 0.002 wt % H₂O) if they are dry. Sample to sample variation is <40% and depends on the humidity during cold pressing. On the basis of spot measurements of a single sample the distribution of waterrelated species is quite homogeneous with <10% difference from average values. We observed no significant difference in water content of samples before or after deformation. No loss of hydrogen by diffusion through the metal jacket during deformation runs could be detected.

FTIR spectra of a few specimens recorded at liquid nitrogen temperature showed a weak ice peak around 3200 cm⁻¹ wavenumber, indicating the presence of freezable water in fluid inclusions [Aines and Rossman, 1984; K. Röller, personal communication, 1999]. Extrapolation of the data of Behrens [1994] suggests that plagioclase may incorporate 0.05 wt % H₂O at T = 1273 K and a water pressure of $P_{H2O} = 300$ MPa. Natural



Figure 1. Representative FTIR absorbance spectra of dry and wet anorthite aggregates samples in the mid-IR range (3 μ m wavelength). Using the maximum peak heights near 3600 cm⁻¹, the calculated water content of the wet sample (An10) is 9800 ppm H/Si before deformation (corresponding to 635 ppm H₂O (wt)) and 8400 ppm H/Si (545 ppm H₂O (wt)) after deformation. The dry undeformed sample (An28) contains \approx 500 ppm H/Si (33 ppm H₂O (wt)). An additional peak at a wavenumber of \approx 3200 cm⁻¹ was found in undeformed wet sample (An15, 10,400 ppm H/Si) when measured at cryogenic temperature. The small sharp peaks at 2850 and 2930 cm⁻¹ result from resin used in sample preparation.

feldspars can contain up to several thousand parts per million H/Si [Kronenberg et al., 1986; Behrens, 1994; Beran, 1987; Hofmeister and Rossman, 1985].

2.3. Microstructures

Figure 2a is an optical micrograph taken from an ultrathin $(<10 \ \mu m)$ section of the wet starting material. Prismatic anorthite grains are common with an average aspect ratio of 2.7. Grain boundaries are facetted, and 120° triple junctions are scarce. Twins with fine-spaced lamellae are abundant. A few larger grains show undulatory extinction. In wet samples some spherulites were found with needle-shaped grains up to 200 µm long and arranged in radial fashion. Spherulites were previously observed to crystallize from undercooled plagioclase melt [Abe et al., 1991; Kirkpatrick et al., 1979; Lofgren, 1974]. Farver and Yund [1995b] found spherulitic texture in hot-pressed orthoclase if the particle size of the starting glass powder was >4 μ m. Since our starting glass powder contains particles <60 µm, it is possible that some of the larger glass particles crystallized as spherulites. Their volume fraction in the wet starting material was 4 ± 3 vol % on average. No spherulites were found in the dry samples.

The arithmetic mean grain size d was determined from samples before and after deformation using the line intercept method [Underwood, 1970]. Polished sections were etched with diluted hexafluorosilicic acid (34% H₂SiF₆) and examined in a scanning electron microscope (SEM Zeiss DSM 962). The mean intercept length L was determined along 12 lines across the section oriented in 15° intervals. L is converted to grain size by multiplying L by a factor of 1.95 and assuming an average aspect ratio of 2.7 and a rectangular prism grain shape [Dimanov et al., 1998]. The mean grain size is $3.4 \pm 0.2 \mu m$ for wet samples and



Figure 2. (a) Optical micrograph of hot-pressed anorthite (An21, crossed nicols, scale bar 12.5 μ m). The ultrathin section photograph reveals fine grains with high aspect ratios of about 2 to 3 that are randomly oriented. The average grain size is $\approx 3.5 \mu$ m. (b) Grain size distribution of sample An23 (wet, undeformed). The arithmetic mean grain size is $3.2 \pm 1.6 \mu$ m. (c) TEM bright-field micrograph of sample An 24 shows twins and sutured grain boundaries. Spacing of the twins is <20 nm. (d) Intracrystalline submicron-sized fluid inclusions (arrows) surrounded by a glass rim are abundant in the wet starting material (sample An12). The trace of small contamination spots at a distance from the central bubble stem from EDX analysis that showed a Si enrichment in the center of the inclusion.

 $2.7 \pm 0.1 \ \mu m$ for dry samples. For some specimens we measured the grain size distribution with the aid of a digital imaging system. It is typically left-skewed (lognormal), as shown in Figure 2b. Since grains smaller than 0.5 $\ \mu m$ diameter could not be resolved from etched SEM sections, we estimated their volume proportion in two samples using transmission electron microscope images (TEM, Philips CM 200). The volume fraction of submicron-sized grains is $\approx 5\%$ after hot pressing at 1323 K for 4 hours and <1 vol % after annealing at 1473 K for 10 hours.

To investigate the grain growth kinetics of wet anorthite aggregates, we hot isostatically pressed four samples between 1323 K and 1473 K for 4 hours and annealed four samples at 1473 K between 4 and 10.5 hours. No significant grain growth occurred under these conditions. However, *Dresen et al.* [1995] found substantial grain growth in anorthite in the presence of water. In their case the samples were crystallized from a glass block. Since the glass powder contains water adsorbed on particle surfaces, grain growth may be inhibited by pinning caused by abundant pores [*Olgaard and Evans*, 1988]. For rigid inclusions the stabilized grain size d_{max} depends on second phase diameter d_s and volume fraction f:

$$d_{\max} = C \left(d_s / f^{\mathrm{m}} \right). \tag{1}$$

C is a constant of ≈ 1 ; m is a measure of second phase dispersion. For a random distribution, particle distribution m =1.0, and if particles lie on corners, m = 1/3 [Olgaard, 1985; Evans et al., 2000]. For a sample porosity of 1% and an average pore size of $d_s = 0.5 \mu m$ with pores preferentially located at grain boundaries (m = 0.5), equation (1) gives a rough estimate of the stabilized grain size of $\approx 5.0 \mu m$.

The dislocation and fluid inclusion density of undeformed and deformed specimens was determined from 300 TEM micrographs [*Underwood*, 1970]. The dislocation density ρ of undeformed anorthite is $\approx 5 \times 10^{11} \text{m}^{-2}$. This is a lower bound, since we did not quench the samples after hot isostatic pressing. Most dislocations are gently curved, suggesting that climb is relatively difficult even at 1473 K. Twinning is common in the undeformed material (Figure 2c). It may result from misfit stresses that accrue during crystallization and ordering of anorthite [*Putnis*, 1992; *Xu et al.*, 1997].

The undeformed wet specimens contain abundant intracrystalline fluid inclusions up to several hundred nanometers in diameter (Figure 2d). As shown in the low-temperature FTIR measurements, the inclusions are partially freezable. The amount of fluid inclusions in wet (dry) material is 0.2 (0.01) vol % on average, equivalent to about 0.07 (0.004) wt % H₂O in anorthite if all inclusions are filled with water. These values agree very well with the hydrogen content determined by FTIR, suggesting that most of the water is incorporated as fluid inclusions during hot isostatic pressing. In both undeformed and deformed wet samples we found <0.7 vol % glass, and in dry samples the glass content was <0.2 vol %. Frequently, the glass surrounding the fluid inclusions forms a shell (Figure 2d); in wet specimens some triple junctions contained glass.

In TEM, no glass was observed along grain boundaries to a resolution of 2 nm. Energy dispersive X-ray analysis (EDX) of the glass in 19 different locations shows a significant Si enrichment with respect to the adjacent anorthite crystal. This Si enrichment may be due to local nonstoichiometric composition of the starting glass powder or due to melting of anorthite under the experimental conditions in the presence of water.

2.4. Mechanical Tests and Data Processing

We performed constant load (creep) tests at 300 MPa confining pressure in a Paterson-type gas apparatus. Temperature was varied from 1093 K to 1493 K. Specimens were jacketed in iron with 0.3 mm wall thickness and nominally drained to the atmosphere. Force was corrected for strength of the iron jacket and converted to axial stress by taking sample distortion into account. True (natural) strain and strain rates were determined from axial displacement and corrected for system compliance. Considering the propagation of errors associated with testing and

data processing, the uncertainty of stress and strain rate data is <3%. Temperatures are accurate to within 0.5%.

3. Results

3.1. Mechanical Data

3.1.1. Stress exponents. About 190 creep tests were performed on 20 wet and 5 dry samples at stresses between 30 to 600 MPa; the resulting steady state strain rates were between $2x10^{-6}$ s⁻¹ and $1x10^{-3}$ s⁻¹ (Table 1). The strain for each step at constant load was $\approx 3\%$, and the total strain accumulated in individual samples was 25% on average. Steady state creep was observed after 1-1.5% strain, but the amount of transient (primary) creep increased at higher loads, probably due to the elongated grain shape [*Deng et al.*, 1999].

Figure 3a shows typical examples of the variation of strain rate with applied stress. The reproducibility of strain rate for a given stress is better than a factor of 2 (e.g., samples An10, An45, and An4). For temperatures below ≈ 1473 K, wet samples are weaker than dry specimens; e.g., at T = 1393 K, wet samples An10, An45, and An4 are weaker than sample An39 (partially dried), which itself is weaker than the dry sample An42 (Figure 3a). The strain rate sensitivity increases with stress from $n \approx 1$ to $n \approx 3$.

In most creep tests the load was increased stepwise at constant temperature. We also performed a few tests at constant load and varied T. Here the resulting strain rates were systematically higher when T was increased and lower when T was decreased, as compared to the respective stress-stepping tests at the same conditions. This may be due to small variations in grain size between samples subjected to different temperature treatments. If the stress was back-stepped at constant T, the resulting strain rate was higher compared to the preceding strain rate at similar load. Possibly, the microstructure of the specimens did not fully equilibrate with stress in these experiments. In addition, we performed a displacement-controlled experiment at constant displacement rates on a wet specimen at T = 1373 K. Strain rates were varied in three steps between 1×10^{-3} , 1×10^{-4} , and 1×10^{-5} s⁻¹. The flow stress-strain rate data from these experiments agree with results from creep tests within the observed sample to sample variation.

The mechanical data were fitted using a multilinear regression to a power law creep equation of the form

$$\dot{\varepsilon} = A \sigma^{n} d^{-m} \exp\left(-Q/RT\right), \qquad (2)$$

where $\dot{\varepsilon}$ is strain rate (s⁻¹), A is a preexponential constant, σ is stress (MPa), d is grain size (μ m), Q is the activation energy (kJ mol⁻¹), T is temperature (K), R is the gas constant, and n and m are the stress exponent and the grain size exponent, respectively.

The results of regression analysis are given in Table 2. For analysis we excluded all data from temperature or stress backstepping experiments. Furthermore, data that appeared to be in the transitional regime (see Figure 3a) and some high stress data that are likely to be in the power law breakdown regime [e.g., *Tsenn and Carter*, 1987] were excluded from regression. Figures 3b and 3c show plots of log stress versus log strain rate. The slope *n*, i.e., the stress exponent, increases from $n \approx 1.0 \pm 0.1$ to $n \approx 3.0 \pm 0.1$ with increasing stress, indicating a transition from diffusion to dislocation creep. The dashed lines represent the fit of the multilinear regression to the data. The boundary between the two creep regimes is different for dry and wet samples.

Samp	le ^b	<i>Т</i> , К	σ, MPa	έ, s ⁻¹	ε, %	Comment
An4	N	1389	32.4	1.76 x 10 ⁻⁵	3.26	OH = 12,520 ppm H/Si
	N ^c	1393	65.2	3.39 x 10 ⁻⁹	2.98	$d = 3.15 \pm 0.27 \mu m$
	N	1393	140.5	1.07 x 10 ⁻⁴	3.25	$\Phi = 0.9\%$
						$\alpha = 0\%$
						axial splitting
An10	N	1393	32.1	1.99 x 10 ⁻⁵	2.73	OH = 9720 ppm H/Si
	NL	1389	52.4	2.58 x 10 ⁻⁵	2.89	$d = 3.26 \pm 0.12$ µm
	N	1392	83.4	3.97 x 10 ⁻⁵	2.69	$\Phi = 1.1\%$
	N/P	1392	184.9	1.12 x 10 ⁻⁴	3.60	$\alpha = 2\%$
	P	1391	290.0	3.01 x 10 ⁻⁴	5.44	stress backstep
	Pc	1391	417.0	1.57 x 10 ⁻³	9.48	F
	Р	1391	300.5	9.12 x 10 ⁻⁴	5 24	
An12	NL	1493	32.2	6.21 x 10 ⁻⁵	3 14	OH = 8510 npm H/Si
AII12	N	1442	31.0	2.21×10^{-5}	3.14	$d = 3.68 \pm 0.22$ µm
	Ň	1341	301	3.08×10^{-6}	171	$\Phi = 1.00 \pm 0.22 \mu m$
	N	1341	62.0	6.04 x 10 ⁻⁶	1 78	$\alpha = 1\%$
	N	1341	146 2	1.44×10^{-5}	1 73	temperature steps
	P	1341	267.8	4.21 x 10 ⁻⁵	2.34	amporature stops
	3.76	1002		0.05 105	A / 7	OU 0200
An15	N	1293	55.4	9.95 x 10"	2.67	OH = 9320 ppm H/Si
	N	1293	31.9	7.88 x 10°	1.13	$a = 3.34 \pm 0.35 \mu\text{m}$
	N	1292	31.2	8.66 x 10 ⁻⁰	0.88	$\Phi = 1.2\%$
	N	1292	53.7	1.19 x 10 ⁻⁵	1.12	$\alpha = 5\%$
	N ^r	1292	82.1	1.58 x 10 ⁻⁵	2.27	stress backstep
	N ^c	1292	134.5	2.58 x 10 ⁻⁵	2.78	
	N° N/D	1292	195.2	4.10 x 10 ⁻⁵	2.63	
	N/P את	1291	292.1	1.1/X IU [*]	2.23	
	r P'	1291	403.I 618 5	1.00×10^{-4}	2.79	
	•	. 271	010.0		5.71	
An18	N	1275	31.7	6.10 x 10 ⁻⁶	1.53	OH = 11,480 ppm H/Si
	N	1273	57.4	1.06 x 10 ⁻⁵	1.72	$d = 3.30 \pm 0.35 \mu \mathrm{m}$
	N ^e	1273	88.7	1.56 x 10 ⁻⁵	2.49	$\Phi = 1.2\%$
	N°	1273	211.9	4.08 x 10 ⁻⁵	1.84	$\alpha = 9\%$
	P ^c	1273	409.6	1.04 x 10 ⁻⁴	2.43	temperature steps
	P	1273	594.5	2.33 x 10 ⁻⁴	3.88	lateral fractures
	P	1191	569.6	5.92 x 10 ⁻⁵	2.97	
	D	1093	557.3	4.84 x 10 ⁻ °	1.66	
An20	Nu	1142	87.7	2.20 x 10 ⁻⁶	1.27	OH = 9740 ppm H/Si
	N ^L	1142	144.7	4.16 x 10 ⁻⁶	1.72	$d = 3.66 \pm 0.37 \mu \text{m}$
	NĽ	1142	218.5	7.03 x 10 ⁻⁶	2.46	$\Phi = 1.0\%$
	NL	1141	422.4	1.29 x 10⁻⁵	2.48	$\alpha = 2\%$
	N/P	1140	619.3	1.98 x 10 ⁻⁵	2.87	temperature steps
	Р	1193	496.4	6.24 x 10 ⁻⁵	2.33	-
	Р	1273	475.2	2.37 x 10 ⁻⁴	3.54	
	Р	1346	447.8	6.08 x 10 ⁻⁴	4.78	
An21	Nu	1347	32.5	9.12 x 10 ⁻⁶	1.89	OH = 10.770 ppm H/Si
	N	1347	53.3	1.38 x 10 ⁻⁵	2.18	$d = 3.47 \pm 0.47$ um
	N°	1346	84.7	2.17 x 10 ⁻⁵	2.10	$\Phi = 0.9\%$
	N/P	1346	212.7	7.06 x 10 ⁻⁵	2.00	$\alpha = 2\%$
	P°	1346	310.4	1.41 x 10 ⁻⁴	3.13	temperature step
	P°	1346	392.8	2.69 x 10 ⁻⁴	4.45	
	P	1346	457.8	5.67 x 10 ⁻⁴	5.50	
	Р	1442	212.0	8.18 x 10 ⁻⁴	5.49	
An77	Nc	1373	80.6	3 15 x 10 ⁻⁵	1 70	OH = 10.710 ppm H/Si
11122	Nº	1373	145.4	5.68×10^{-5}	2 33	d = 3.26 + 0.43 um
	 N/P	1372	217.5	1.13×10^{-4}	2.15	$\Phi = 0.9\%$
	P	1372	3157	2.55×10^{-4}	3.70	$\alpha = 3\%$
	Р ^с	1372	394.8	5.55×10^{-4}	5.70	stress backsten
	P	1373	451.6	1.74×10^{-3}	5.47	few axial fractures
	N	1373	42.3	5.02 x 10 ⁻⁵	3.00	
4=00	A/C	1474	20 5	6 40 10-5	1.05	OU - 0670 11/8:
A023	IV N ^C	14/4	54.5 55 2	0.49 X 10 ⁻¹	1.90	$d = 3.07 \pm 0.45$ mm
	N/D	14/3	33.3 130 0	1.01×10 2 04 v 10 ⁻⁴	2.31	a = 3.27 ± 0.43 μm Φ = 0.9%
	NL	1473	۶1 ج ۱	2.74 x 10 1 74 x 10 ⁻⁴	3.27	$\alpha = 3\%$
	N°	1472	100 8	2.07×10^{-4}	J.JZ 275	stress backsten
		1111	100.0	6.07 A 10	4.13	saless sacratep

Table 1. Mechanical Data for Creep of Synthetic Anorthite^a

Sample ^b		<i>T</i> , K	σ, MPa	έ, s ^{.1}	Е, %	Comment
	N/P	1470	148.4	3 39 x 10 ⁻⁴	3 70	
	p	1470	176.1	5.05×10^{-4}	4 70	
	P	1470	218.5	9.67×10^{-4}	6.23	
	-		21012	1.10 10-4	3 22	OH 0220 H/G
An24	N ²	1441	84 9	1.12×10^{-4}	3.32	OH = 9230 ppm H/S1
	N/P	1439	133.3	2.06 X 10 ⁻⁴	3.57	$a = 2.94 \pm 0.41 \mu m$
	N/P	1437	101.7	2.55×10^{-4}	3.50	$\Phi = 1.0\%$
	Г D	1437	100.0	3.29 X 10 4.06 x 10 ⁻⁴	3.00	u = 9%
	Г D ^L	1457	217.0	4.00×10^{-4}	4.02	muai lengui = 17 mm
	Г D ^L	1430	217.0	7.80 × 10 ⁻⁴	6.81	
	P	1437	234.3	1.09×10^{-3}	8 3/	
	1	1437	270.1	1.70 x 10	0.54	
An25	N	1442	55.0	7.78 x 10 ⁻⁵	2.05	OH = 8500 ppm H/Si
	N	1442	85.0	1.15 x 10 ⁻⁴	2.06	$d = 3.32 \pm 0.45 \mu m$
	N/P	1442	134.8	1.94 x 10 ⁻⁴	3.03	$\Phi = 1.0\%$
	P	1441	199.5	3.76 x 10 ⁻⁴	3.62	$\alpha = 9\%$
	P	1441	215.9	5.09 x 10 ⁻⁴	4.41	
	Pu	1440	236.8	6.37 x 10 ⁻⁴	5.14	
	P	1439	239.9	8.13 x 10 ⁻⁴	7.01	
	Р	1440	255.6	1.18 x 10 ⁻³	5.74	
		1217	210.2	6 00 10-5	2.04	011 0((0 11/0)
An27	N/P	1317	318.3	6.92 x 10 ⁻⁴	2.04	OH = 8000 ppm H/S1
	P ^c	1317	353.1	1.12×10^{-4}	2.33	$d = 3.50 \pm 0.35 \mu m$
	P*	1317	394.7	1.65 x 10 ⁻⁴	3.01	$\Phi = 2.5\%$
	P -	1317	420.2	2.32×10^{-4}	3.41	$\alpha = 12\%$
	P' ni	1316	440.2	3.00 X 10 ⁻⁴	3.74	
	P	1317	506.7	5.50 X 10	4.10	
Ап28	N	1468	48.6	1.12 x 10 ⁻⁴	2.02	OH = 500 ppm H/Si
	N°	1467	58.8	1.23×10^{-4}	2.12	$d = 2.72 \pm 0.41 \mu\text{m}$
	NL	1468	90.0	1.76×10^{-4}	2.22	$\Phi = 0.9\%$
	N	1468	102.4	1.94 x 10 ⁻⁴	2.17	$\alpha = 0\%$
	NL	1467	157.0	3.03 x 10 ⁻⁴	2.60	
	P	1467	180.0	3.65 x 10 ⁻⁴	3.23	
	Pu	1467	197.7	4.57 x 10 ⁻⁴	3.82	
	Pu	1467	211.8	5.41 x 10 ⁻⁴	4.35	
	P	1467	245.1	7.94 x 10 ⁻⁴	5.13	
An29	N	1368	61.6	6 95 x 10 ⁻⁶	2.02	OH = 460 ppm H/Si
	N	1368	83.3	8.69 x 10 ⁻⁶	1.76	$d = 2.72 \pm 0.51 \mu\text{m}$
	N ^c	1367	157.0	1.62 x 10 ⁻⁵	2.17	$\Phi = 0.9\%$
	N	1365	232.1	2.69 x 10 ⁻⁵	2.40	$\alpha = 0\%$
	PL	1366	400.3	6.82 x 10 ⁻⁵	2.42	
	P	1366	428.2	9.53 x 10 ⁻⁵	2.34	
	P'	1366	477.0	1.55 x 10 ⁻⁴	1.82	
4 21	N 7L	1000	105.5	7 14 - 10.0	214	OU 12 260 U/C
Ansi	N° NL	1223	105.5	7.14 x 10°	2.14	OH = 12,360 ppm H/S
	N-	1222	210.7	2.17×10^{-5}	2.44	$a = 3.42 \pm 0.05 \mu\text{m}$
		1222	275.5	5.51 X 10 4.69 10 ⁻⁵	2.25	$\Psi = 0.9\%$
	/\/r D	1222	320.1	4.06 X 10 6 52 x 10 ⁻⁵	2.92	$\alpha = 1270$
	Г D	1220	377.1	0.32×10^{-5}	2.33	micromactures
	r D	1220	433.0	7.06 X 10 9.91 v 10 ⁻⁵	2.57	
	Г Р	1220	595 7	9.21 x 10 ⁻⁵	2.50	
	-					
An32	N ^c	1421	86.1	7.02 x 10 ⁻⁵	2.21	OH = 12,210 ppm H/S1
	N	1420	96.5	7.89 x 10 ⁻⁵	1.93	$d = 3.01 \pm 0.10 \mu m$
	N	1420	127.2	1.09 x 10 ⁻⁴	2.24	$\Phi = 0.8\%$
	N/P	1420	152.9	1.57 x 10 ⁻⁴	2.43	$\alpha = 10\%$
	P	1420	189.5	2.13 x 10 ⁻⁴	2.68	
	P	1419	212.6	2.92 x 10 ⁻⁴	3.03	
	P	1419	216.2	3.68 x 10 ⁻⁴	2.57	
	P	1418	247.7	4.83 x 10 ⁻⁴	3.12	
	P	1418	284 7	7.35 x 10 ⁻⁴	4.75	
	Р	1419	289.3	1.08 x 10 ⁻³	4.73	
An33	N ^c	1374	110.5	3.55 x 10 ⁻⁵	2.09	OH = 14.830 ppm H/Si
	N°	1373	107.3	2.78 x 10 ⁻⁵	2.57	$d = 3.18 \pm 0.58 \mu\text{m}$
	N°	1374	159.1	4.64 x 10 ⁻⁵	3.01	$\Phi = 1.1\%$
	N/P	1373	189.4	7.13 x 10 ⁻⁵	2.88	$\alpha = 70\%$
	N/P	1373	243.7	1.14 x 10 ⁻⁴	2.56	

Table 1. (continued)

Table 1. (continued)

Samp	Sample ^b T,		σ, MPa	ė, s ⁻¹	ε, %	Comment
	P°	1371	264.5	1.74 x 10 ⁻⁴	3.03	
	P	1372	280.4	2.21 x 10 ⁻⁴	3.45	
	P	1372	300.7	3.10 x 10 ⁻⁴	3.96	
	P	1372	322.6	4.67×10^{-4}	4.35	
	Ρ	1371	319.6	9.37 x 10⁴	7.17	
An35	N	1372	131.2	6.63 x 10 ⁻⁵	2.05	OH = 14,850 ppm H/Si
	N	1373	127.3	6.43 x 10 ⁻⁵	2.01	$d = 3.59 \pm 0.17 \mu m$
	N	1372	157.8	8.57 x 10 ⁻⁵	2.63	$\Phi = 0.9\%$
	N/P	1373	208.8	1.98 x 10⁴	1.45	$\alpha = 40\%$
An36	N	1101	121.1	1.09 - 10-6	1.00	OU 4200 U/0
Alibu	N	1191	131.1	1.08 X 10 ⁻⁶	1.23	$d = 2.82 \pm 0.12$
	N	1190	348.8	1.02 x 10 1.91 x 10 ⁻⁶	1.40	$a = 3.85 \pm 0.15 \mu m$
	Nu	1190	479.8	2.73 x 10 ⁻⁶	1.74	$\alpha = 0\%$ dried
						a = 070, and
An39	N ^c	1387	87.5	2.28 x 10 ⁻⁵	2.58	OH = 7350 ppm H/Si
	N ^c	1386	134.9	3.45 x 10 ⁻⁵	2 47	$d = 3.48 \pm 0.59 \mu \mathrm{m}$
	N/P	1388	199.3	6.14 x 10 ⁻⁵	3.22	$\Phi = 0.7\%$
	P ^c	1388	264.3	1.13 x 10 ⁻⁴	3.41	$\alpha = 0\%$, dried
	ף. זע	1388	294.6	1.56 x 10 ⁻⁴	4.23	
	Г D ^C	1300	319.3	2.23×10^{-4}	4.22	
	1	1366	507.4	5 42 X 10	5.58	
An41	N	1430	50.0	3.70 x 10 ⁻⁵	1.90	OH = 710 nnm H/Si
	N ^c	1429	72.9	4.38 x 10 ⁻⁵	2.52	$d = 2.55 \pm 0.21$ µm
	N^{c}	1430	97.2	5.74 x 10 ⁻⁵	2.62	$\Phi = 1.0\%$
	N ^c	1429	148.4	8.46 x 10 ⁻⁵	3.20	$\alpha = 0\%$
	Ρ	1428	236.4	1.64 x 10 ⁻⁴	4.06	
	P	1429	261.4	2.28 x 10 ⁻⁴	4.30	
	P ^c	1429	289.1	2.86×10^{-4}	4.67	
	P° Dí	1429	331.2	4.12 x 10 ⁻⁴	5.66	
	P^{*}	1429	355.5	5.95 x 10 ⁻⁴	6.06	
An42	N	1389	77.2	1.13 x 10 ⁻⁵	2.07	OH = 450 ppm H/Si
/ _	Nu	1391	194.4	2.98×10^{-5}	3 14	$d = 2.61 \pm 0.19 \mu m$
	Nu	1390	246 9	3.76 x 10 ⁻⁵	3.20	$\Phi = 0.9\%$
	P	1390	312.5	5.70 x 10 ⁻⁵	3.91	$\alpha = 0\%$
	P	1390	331.5	7.24 x 10 ⁻⁵	3.97	
	P	1390	366.5	9.80 x 10 ⁻⁵	4.34	
	P ^c	1390	424.1	1.48×10^{-4}	3.65	
	<i>P</i> *	1390	453.6	1.98 x 10 *	4.23	
An43	N	1253	77 3	5 96 x 10 ⁻⁶	1.61	OH - 19 250 ppm H/St
	N	1253	122.5	1.36×10^{-5}	1.79	$d = 3.93 \pm 0.23$ µm
	NL	1250	169.4	1.88 x 10 ⁻⁵	2.59	$\Phi = 1.6\%$
	NĽ	1251	245.9	3.15 x 10 ⁻⁵	3.21	$\alpha = 90\%$
	Ν	1251	314.4	4.38 x 10 ⁻⁵	2.75	microfractures
	N	1251	347.1	4.94 x 10 ⁻⁵	3.16	
	N/P	1245	402.9	5.30 x 10 ⁻⁵	3.48	
An44	N	1480	36.6	0 30 × 10-5	2 51	
	N	1430	51.3	9.30×10^{-4}	2.31	$d = 2.80 \pm 0.38 \text{ µm}$
	N	1479	717	1.14×10^{-4}	3 20	$a = 2.69 \pm 0.58 \mu m$
	N°	1478	105.2	1.80×10^{-4}	3.57	$\alpha = 1\%$
	N°	1478	105.5	1.98 x 10 ⁻⁴	2.86	
	N ^c	1480	154.4	2.77 x 10 ⁻⁴	4.05	
	P	1479	177.7	3.68 x 10 ⁻⁴	4.60	
	P	1479	197.6	4.87 x 10 ⁻⁴	3.85	
	P°	1479	225 8	6.48 x 10 ⁻⁴	4.68	
An45	N	1392	47 5	2 50 x 10 ⁻⁵	2 16	OH - 16 220 mm H/S
, m+5	NL	1392	53.0	2.50×10^{-5}	2.10	$d = 3.45 \pm 0.30$ mm
	N ^c	1391	76.2	3.42×10^{-5}	2.49	α – 5.45 ± 0.50 μm Φ = 1.1%
	NL	1391	122.0	6.07 x 10 ⁻⁵	3.18	$\alpha = 90\%$
	N/P	1391	178.0	1 17 x 10 ⁻⁴	3.30	
	P	1390	216.6	1.88 x 10 ⁻⁴	4.05	
	P	1390	245.0	2.82 x 10 ⁻⁴	4.70	
	P^{ν}	1390	253.2	3.55 x 10 ⁻⁴	5.69	
	r	1391	290.4	0.06 x 10 ⁻⁴	6.61	

^a T is temperature, σ is flow stress, $\dot{\varepsilon}$ is (natural) strain rate, ε is strain, OH is water content determined by FTIR, d is mean grain size, ϕ is porosity, and α is vol % of spherulitic grains. ^b Index N and P indicate diffusion and dislocation creep regime, respectively. ^c Data are used for data fit to a power law.



Figure 3. Log stress versus log strain rate plots for deformed anorthite aggregates. (a) Strain rate which increases with increasing stress, temperature, and water content over the investigated range of conditions. (b) and (c) Data for all dry and wet samples are separated into two deformation regimes with stress exponent n = 1 and n = 3, indicating diffusion creep and dislocation creep, respectively. The boundary between the two regimes is shifted toward higher stresses with decreasing temperature and water content. Dashed lines are best-fit regression lines. Experimental error is indicated for one data point.

Regime		n	Q, kJ mol ⁻¹	$\log A^{\mathrm{b}}$, MPa ⁻ⁿ µm ^m s ⁻¹	m ^b
Diffusion Creep	Dry Wet	1.0 ± 0.1 1.0 ± 0.1	467 ± 16 170 ± 6	12.1 ± 0.6 1.7 ± 0.2	3 3
Dislocation Creep	Dry Wet	3.0 ± 0.4 3.0 ± 0.2	648 ± 20 356 ± 9	12.7 ± 0.8 2 6 ± 0.3	0 0

 Table 2. Power Law Creep Parameters for Diffusion and Dislocation Creep of Dry and Wet

 Synthetic Anorthite^a

^a n is stress exponent, Q is activation energy, A is preexponential factor, and m is grain size exponent. ^b For determination of A the grain size exponent was set to m = 0 for dislocation creep, and m = 3 for diffusion creep (see text for discussion) It should be noted that given uncertainties are standard deviations of multiple regression analysis and do not consider experimental accuracy.



Figure 4. Arrhenius diagrams for (a) diffusion and (b) dislocation creep of dry and wet specimens, normalized to 10 MPa differential stress. The activation energy for dry samples is ≈ 300 kJ mol⁻¹ higher than that for wet specimens in both regimes. Experimental error is indicated for one data point.

3.1.2. Grain size sensitivity. Since we were not able to measure the grain size exponent m explicitly, we set m = 0 for dislocation creep. For diffusion creep, m was assumed to be 3. Wang et al. [1996], Dimanov et al. [1999], and Xiao [1999] reported m values from 2.5 to 2.9 \pm 0.3, implying grain boundary diffusion-controlled creep for similar material in uniaxial and triaxial creep experiments.

3.1.3. Activation energy. The activation energy Q for diffusion creep of anorthite varies from 170 ± 6 kJ mol⁻¹ for wet aggregates to 467 ± 16 kJ mol⁻¹ for dry samples (Figure 4a). For dislocation creep we found activation energies of 356 ± 9 and 648 ± 20 kJ mol⁻¹ for wet and dry samples, respectively (Figure 4b). The ratios of activation energies for diffusion and dislocation creep are between 1.4 for dry samples and 2.1 for wet specimens. This is in agreement with activation energy ratios commonly observed for volume to grain boundary diffusion-controlled creep ranging from 1.5 to 2 [e.g., Schmid et al., 1977].

Two samples (An36 and An39) were dried but still retained 4500-7500 ppm H/Si (Table 1). Regression analysis of the few creep data on these samples yields n = 1 and $Q = 268 \pm 1$ kJ mol⁻¹ in the diffusion creep regime, which is between the activation energies of dry and wet samples (Table 3).

3.2. Microstructure of Deformed Samples

In almost all experiments the load was increased in steps until samples were deformed in the dislocation creep field. In SEM, no strong shape-preferred orientation of the grains was observed, and no marked lattice-preferred orientation was seen in optical microscopy using a gypsum plate. The deformed synthetic anorthite aggregates show more submicron-sized recrystallized grains and less big twins than the starting material. Occasionally,

the smaller dislocation-free recrystallized grains show dihedral angles close to 120°, especially at high temperature and in dry samples. However, larger acicular or tabular grains containing few deformation twins are more common. Grain boundaries of the original grains are strongly sutured. Grains with high dislocation density and twin lamellae appear to be consumed by less deformed grains (Figure 5a). The dislocations are inhomogeneously distributed and very high dislocation densities are found within individual grains. Dislocations show curved and straight segments. The dislocation patterns are variable with free dislocation segments and some dislocation walls (Figure 5b). Dislocation networks are scarce but more frequent at higher temperatures. The average dislocation density is $\rho \approx 1.4 \times 10^{12} \text{ m}^{-2}$, i.e., ≈ 3 times higher than in the starting material. However, for technical reasons, samples could not be quenched under load, so the dislocation density represents a lower bound.

In deformed specimens, fluid inclusions decorate grain boundaries. Intracrystalline fluid inclusions are commonly surrounded by glass. More triple junctions are filled with Sienriched melt in deformed samples than in the starting material (Figure 5c). The amount of intracrystalline glass is small (<0.2 vol %) in dry samples and decreases with increasing annealing temperature in dry specimens. At T > 1323 K, wet samples contain more glass than at lower temperatures. In a few cases we observed melt on two-grain boundaries in samples deformed at high temperature.

4. Discussion

The log stress-log strain rate plots in Figure 3 indicate a low stress sensitivity of the strain rate for dry and wet anorthite

 Table 3. Compilation of Power Law Creep Parameters for Diffusion and Dislocation Creep of Synthetic Silicate

 Aggregates^a

No,	Rock	log A	Q	n	m	H/Si, ppm	Remark	Reference					
	Dislocation Creep Regime												
1	An_{100}	12.7 ± 0.6	648 ± 20	3	0	640 ± 260	dry	this study					
2	An ₁₀₀	2.6 ± 0.3	356 ± 9	3	0	11500 ± 2900	wet	this study					
3	01	3.28	420	3	0	1500 - 3000	wet, Q estimated	Ka86					
4	Qtz	-7.18	135	31	0	1000 - 12600	wet, estimation of parameters	Pa90					
	Diffusion Creep Regime												
5	Ol	4 83	290	1	2	1 0	dry, Q, m estimated	Ka86					
6	01	618	250	1	3		wet, Q, m estimated	Ka86					
7	An ₁₀₀	12.1 ± 0.6	467 ± 16	1	3	640 ± 260	dry	this study					
8	An ₁₀₀	1.7 ± 0.2	170 ± 6	1	3	11500 ± 2900	wet	this study					
9	An ₁₀₀	1.5 ± 0.3	162 ± 8	1	3	11500 ± 2900	wet, parameter fit for $T < 1343$ K	this study					
10	An_{100}	5.3 ± 0.5	267 ± 13	1	3	11500 ± 2900	wet, parameter fit for $T \ge 1343$ K	this study					
11	An_{100}	5.1 ± 0.1	268 ± 1	1	3	5870 ± 2090	dried samples An36 and An39	this study					
12	An_{100}	1.5 ± 1.2	170 ± 29	1	2.8	8500 ± 1500	wet, predried, (refitted)	Xi99					
13	An ₁₀₀	1.4 ± 0.4	151 ± 40	1	2.8	22000 ± 8000	wet, as received, (refitted)	X199					
14	An_{100}	15.8 ± 0.6	585 ± 45	I	2.9	430 ± 120	dry (HUP), uniaxial	Di99					
15	An_{100}	8.6 ± 0.6	377 ± 38	1	2.5	5530 ± 2470	wet (HIP), uniaxial	Di99					
16	An ₁₀₀	9.0 ± 0.8	388 ± 22	1	2.5	3800 ± 750	wet (HIP1), uniaxial, (refitted)	D199					
17	An ₁₀₀	8.1 ± 0.5	361 ± 13	1	2.5	7000 ± 1000	wet (HIP2), uniaxial, (refitted)	D199					
18	An ₁₀₀	10.6	420 ± 38	1.3	2.5		wet, uniaxial	Wa96					
19	An ₆₀	7.2	338 ± 27	1	2.7		wet, uniaxial, partial melt	D198					

^an is stress exponent, Q is activation energy (kJ mol⁻¹), A is preexponential factor (MPa⁻ⁿ μ m^m s⁻¹), m is grain size exponent, and T is temperature. The data indicated with "uniaxial" are derived at 1 atm conditions, all others are derived at 300 MPa confining pressure. References are 1-2 and 7-11, anorthite (An₁₀₀), this study; 3, 5-6, olivine (Ol), Ka86 [Karato et al., 1986]; 4, quartz (Qtz), Pa90 [*Paterson and Luan*, 1990]; 12-13, anorthite (An₁₀₀), Xi99 [Xiao, 1999] (the flow law parameters are refitted using multiple linear regression analysis); 14-17, anorthite (An₁₀₀), Di99 [*Dimanov et al.*, 1999] (14 are hot uniaxially pressed, HUP, specimens, 16 and 17 are separate refits for hot isostatically pressed, HIP, samples, subdivided into HIP1 and HIP2 samples with slightly different water content); 18, anorthite (An₁₀₀), Wa96 [*Wang et al.*, 1996]; 19, labradorite (An₆₀), Di98 [*Dimanov et al.*, 1998].



b





Figure 5. TEM bright-field microphotographs of deformed anorthite aggregates. (a) Grain boundary migration of a strain-free recrystallized grain toward grains with very high (tangled) dislocation density (arrow, sample An12). (b) Dislocations which are often gently curved or straight sometimes form walls or networks (arrows, sample An18). (c) Fluid inclusions surrounded by glass located in triple junctions or are intracrystalline (arrows, sample An4).

aggregates at low stresses with n = 1 and a transition to a higher stress sensitivity of n = 3 at high stresses. A transition between the dislocation and diffusion creep field was also observed in other studies of fine-grained synthetic aggregates, i.e., olivine [Hirth and Kohlstedt, 1995a, 1995b; Karato et al., 1986], feldspar [Dimanov et al., 1998; Ji and Mainprice, 1986; Meyer et al., 1993; Tullis and Yund, 1991; Xiao et al., 1996], calcite [Walker et al., 1990], anhydrite [Dell'Angelo and Olgaard, 1995], and natural granitic aggregates [Dell'Angelo et al., 1987]. The stress at the transition between regimes decreases with increasing temperatures (Figure 3a).

Raj and Ghosh [1981] and Freeman and Ferguson [1986] suggested that the transition between dislocation and diffusion creep would depend on the grain size distribution of the deformed aggregate; a broad grain size distribution results in a broad transitional regime. Wang [1994] found that the stress exponent may vary considerably due to a small variation in grain size around the critical size for which both mechanisms contribute to the overall strain rate. The grain size distribution of our starting material (Figure 2b) and the bimodal grain size distribution due to dynamic recrystallization probably both contributed to the observed transition between regimes.

4.1. Dislocation Creep Regime

At $T \approx 1273$ K the stresses required to deform the samples at strain rates between $\approx 1 \times 10^{-4}$ and 1×10^{-5} s⁻¹ in the dislocation creep field exceeded the confining pressure by a factor of ≈ 2 . For stresses exceeding the confining pressure, Goetze's criterion [Evans and Kohlstedt, 1995] predicts the onset of dilatant microcracking. Also, at high stresses, power law creep breaks down and the stress exponent increases with stress. Power law breakdown stresses σ_{PLB} are proportional to the shear modulus μ ; for metals and ceramics, $\sigma_{PLB} \approx 10^{-3} \mu$ [Frost and Ashby, 1982] but may be as high as $\approx 10^{-2} \mu$ for silicates [Tsenn and Carter, 1987]. This suggests that the power law breakdown stress for anorthite aggregates σ_{PLB} is $\approx 200-250$ MPa. However, we did not observe a nonlinear increase of n with stress up to stresses of ≈500 MPa. Only in some experiments at the highest stresses did we observe accelerated creep rates due to cracking. Since we did not measure volumetric strain, it is possible that cracking occurred at the highest stresses but remained undetected.

The slightly curved dislocation segments commonly found in the deformed anorthite samples suggest that climb is relatively difficult. Strongly sutured grain boundaries, bulging of grains, and numerous small recrystallized grains suggest that grain boundary migration recrystallization is common [Tullis, 1983; Tullis and Yund, 1985]. In quartzite, Hirth and Tullis [1992] identified three different dislocation creep regimes characterized by different recrystallization mechanisms. At low temperatures, Hirth and Tullis [1992] describe dislocation creep accommodated by grain boundary migration recrystallization (regime I). The microstructures of our deformed wet and dry anorthite samples are similar to the microstructures described by Hirth and Tullis [1992] and to feldspar microstructures recently described by Post and Tullis [1999] typical for low-temperature dislocation creep. However, in our anorthite samples deformed at elevated temperatures, subgrain boundaries were more frequently observed than in previous studies. This suggests an increasing amount of recovery-accommodated creep at high T. The slight increase in melt content in our experiments should not contribute significantly to the deformation rate since it does not wet the grain boundaries and the melt content is <1 vol % [Hirth and Kohlstedt, 1995a; Dimanov et al., 1998].

4.1.1. Activation energy. We found an activation energy of $648 \pm 20 \text{ kJ} \text{ mol}^{-1}$ for dislocation creep of dry samples, whereas in the presence of water it was reduced significantly to $356 \pm 9 \text{ kJ} \text{ mol}^{-1}$. The activation energy for creep at high temperatures $T > 0.5T_m$ (T_m is melting temperature) has been suggested to be similar to that of self-diffusion [*Takeuchi and Argon*, 1976]. However, when dislocation climb is difficult and recovery is slow, dynamic recrystallization will contribute more significantly to the release of stored strain energy [*Cahn*, 1983], and the observed activation energy may reflect a mixed mechanism.

Although the process limiting the creep rate in plagioclase is not clear, it probably involves breaking of Si-O and Al-O bonds in the feldspar lattice. Carpenter [1991] studied the kinetics of Si-Al exchange in synthetic dry anorthite aggregates at 0.1 MPa pressure using antiphase domain coarsening. He obtained activation energies ranging from 506 to 637 kJ mol⁻¹, depending on the growth exponent. Recently, Liu and Yund [1992] NaSi-CaAl interdiffusion investigated by lamellar homogenization. In the presence of 0.1 wt % water, at 1.5 GPa pressure and temperatures <1248 K they found activation energies of 371 and 303 kJ mol⁻¹ for homogenization in Huttenlocher plagioclase and peristerite, respectively. Baschek and Johannes [1995] also investigated the homogenization of a peristerite at 1 GPa pressure and an H₂O mole fraction of 0.5. They obtained an activation energy of 465 kJ mol⁻¹. Grove et al. [1984] determined a Q of 517 kJ mol⁻¹ for homogenization of Huttenlocher lamellae in dry plagioclase at 0.1 MPa pressure.

These activation energies for Si-Al exchange indicate that the kinetics of NaSi-CaAl interdiffusion are strongly affected by the presence of water. They are in the range of values that we determined for dislocation creep, suggesting that Si or Al diffusion may control creep in the high-stress regime. This conclusion is also consistent with the observation that the activation energies determined for lattice diffusion of O and Ca are substantially lower. For oxygen self-diffusion in anorthite the values range from 162 ± 36 kJ mol⁻¹ [*Ryerson and McKeegan*, 1994] to 236 ± 8 kJ mol⁻¹ [*Elphick et al.*, 1988] under dry conditions. An activation energy Q of 110 ± 5 kJ mol⁻¹ was found under wet conditions [*Giletti et al.*, 1978]. For Ca lattice diffusion, *Behrens et al.* [1990] obtained 313 kJ mol⁻¹ in dry An₆₀.

4.1.2. Comparison with other deformation studies. Experimentally derived flow laws for dislocation creep of plagioclase are scarce. Montardi and Mainprice [1986] found a stress exponent of $n \approx 3.9$ and a very high Q of 1332 ± 80 kJ mol⁻¹ for uniaxial creep of dry synthetic anorthite samples with $\approx 50 \ \mu m$ grain size, which was possibly influenced by melting. Meyer et al. [1993] performed uniaxial creep tests on dry synthetic anorthite aggregates with $\approx 3 \ \mu m$ average grain size. They determined $n \approx$ 2.5, suggesting that the samples were deformed in the transitional regime with $Q = 660 \pm 60$ kJ mol⁻¹. Dimanov et al. [1998] deformed fine-grained synthetic An₆₀ with 0.1 wt % H₂O in uniaxial compression. At the transition between the diffusion to dislocation creep field they reported Q = 670 kJ mol⁻¹ and $n \approx 2$ -3. Ji and Mainprice [1987] deformed albite aggregates at 0.1 MPa and found similar strength for "as-received" and predried samples. For temperatures below 1333 K they obtained a stress exponent n = 2.1 and an activation energy of 524 ± 23 kJ mol⁻¹.

With the exception of *Montardi and Mainprice* [1986] the reported activation energies are close to the value of $Q = 648 \pm 20$

26,029

kJ mol⁻¹ that we found for dislocation creep of dry anorthite aggregates. However, composition, water content, and grain size varied between the studies. The activation energy for dislocation creep of our samples with trace amounts of water and deformed at 300 MPa confining pressure is much lower than that for dry samples, but it is also lower than that for samples deformed at atmospheric pressure in the presence of water. The significant reduction of Q suggests that dislocation creep is strongly influenced by water fugacity.

Shelton and Tullis [1981] and Shelton [1981] deformed natural anorthosite and albite rocks in a solid-medium apparatus at temperatures between 973 K and 1373 K and 1.5 GPa confining pressure. For stresses between 200 and 600 MPa they found n = 3.2 and Q = 239 kJ mol⁻¹ for anorthosite. The stress exponent was found to be stress-dependent [Shelton, 1981], and steady state was probably not reached in these experiments [Tullis, 1990]. For albite, Shelton and Tullis [1981] reported n =3.9 and Q = 235 kJ mol⁻¹. Deformation of albite samples dried at 873 K gave a stress exponent of ≈ 4 and Q = 431 kJ mol⁻¹ [Shelton, 1981]. The reported activation energies are significantly lower than for the dry and wet synthetic anorthite samples investigated in this study. The reason for this discrepancy is not clear but may be partly related to a substantially higher water fugacity in the experiments of Shelton and Tullis [1981].

The existing mechanical data for dislocation creep of plagioclase are compiled in an Arrhenius diagram (Figure 6). Two sets of curves can be distinguished: (1) dry Ab_{94} is obtained at 0.1 MPa [*Ji and Mainprice*, 1987], and our dry and wet An_{100} data intersect at ≈ 1500 K. The dry albite is stronger than dry anorthite at low *T*, which may be related to different water fugacity. (2) In comparison, *Shelton's* [1981] data for wet An_{75} ,

and wet and dry Ab₉₆, measured at 1.5 GPa pressure, plot 2 orders of magnitude lower in strain rate at $T \approx 1500$ K. The reason for this discrepancy may be again a difference in water fugacity. Owing to the low activation energy *Shelton's* [1981] data predict faster strain rates for deformation of plagioclase at low temperatures compared to our results and those of *Ji and Mainprice* [1987].

4.2. Diffusion Creep Regime

In the low-stress regime we determined a stress exponent n = 1 for wet and dry samples. Since grain growth was slow in our experiments, we did not determine grain size sensitivity of the creep rate. As already mentioned, the grain size sensitivity is assumed to be m = 3 based on the findings of *Dimanov et al.* [1999], *Wang et al.* [1996], and *Xiao* [1999] on similar material with comparable grain size and range of stresses. An m = 3 suggests that grain boundary diffusion-controlled creep operates in fine-grained anorthite at low stresses. *Tullis et al.* [1996] and *Tullis and Yund* [1991] also described grain boundary diffusion creep in triaxially deformed fine-grained feldspar. At 600-1200 MPa confining pressure and 1173 K temperature with ≈ 0.7 wt % water added, synthetic feldspar aggregates were deformed by diffusion creep. However, in drier samples (≈ 0.2 wt % H₂O), dislocation creep dominated.

4.2.1. Comparison with other studies on feldspar deformation. We found an activation energy of $467 \pm 16 \text{ kJ mol}^{-1}$ for grain boundary diffusion creep of dry samples. The activation energy for diffusion creep of wet samples deformed in the temperature range of 1153 K to 1493 K is $170 \pm 6 \text{ kJ mol}^{-1}$. However, the data may be subdivided into a high- and a low-temperature regime at ≈ 1343 K (Figure 4a). A separate regression



Figure 6. Arrhenius plot of dislocation creep of dry and wet plagioclase, normalized to 100 MPa flow stress. Dry (1) and wet (4) An₁₀₀; confining pressure P = 300 MPa; wet (2) An₇₅; as-received material, P = 1.5 GPa, n = 3.2, Q = 239 kJ mol⁻¹, log A = -3.7 MPa⁻ⁿ s⁻¹; wet (3) Ab₉₆; as-received, P = 1.5 GPa, n = 3.9, Q = 235 kJ mol⁻¹, log A = -5.6 MPa⁻ⁿ s⁻¹; dry (5) Ab₉₆; samples dried at 873 K, P = 1.5 GPa, n = 4, Q = 431 kJ mol⁻¹, log A = 0.9 MPa⁻ⁿ s⁻¹; dry (6) Ab₉₆; as-received material and specimens dried at 1303 K, P = 0.1 MPa, n = 2.1, Q = 524 kJ mol⁻¹, log A = 11.2 MPa⁻ⁿ s⁻¹. Except for our aggregates all samples were obtained from natural rocks. Note that *Carter and Tsenn* [1987] refitted the data for wet Ab₉₆ using multiple linear regression analysis. They found n = 2.8, Q = 245 kJ mol⁻¹, and log A = -2.3 MPa⁻ⁿ s⁻¹.



Figure 7. Arrhenius diagram of grain boundary diffusion creep of synthetic anorthite aggregates, calculated for 10 MPa stress and 5 μ m grain size. Dry samples contain a few hundred parts per million H/Si and wet material several thousand parts per million H/Si. Uniaxial data were obtained at atmospheric pressure, and triaxial data were obtained at 300 MPa confining pressure. Flow law parameters for uniaxially deformed samples are given in Table 3 (14 and 15).

for the data above and below temperatures of 1343 K yields activation energies of 267 ± 13 and 162 ± 8 kJ mol⁻¹ for high- and low-temperature diffusion creep, respectively (Table 3). We found no microstructural evidence for a change in deformation mechanism with temperature. At elevated temperatures some melting may have occurred in the wet samples, leading to a weakening and a higher activation energy [Hirth and Kohlstedt, 1995a]. We therefore consider an activation energy of $170 \pm 6 \text{ kJ}$ mol⁻¹ based on the whole temperature range as a lower bound for diffusion creep of wet anorthite aggregates. This value is significantly lower than the activation energies reported from deformation experiments on plagioclase aggregates performed at atmospheric pressure in the diffusion creep regime [Dimanov et al., 1998, 1999; Ji and Mainprice, 1986; Mercer and Chokshi, 1993; Meyer et al., 1993; Wang et al., 1996; Montardi and Mainprice, 1986]. Recently, Dimanov et al. [1999] compiled diffusion creep data for anorthite aggregates. The activation energies from these studies range from 590 to 770 kJ mol⁻¹ for dry and from 370 to 420 kJ mol⁻¹ for wet anorthite samples, respectively.

In Figure 7 we compare our data with those of *Dimanov et al.* [1999]. At T < 1500 K all samples deformed at atmospheric pressure are considerably stronger than when deformed at 300 MPa confining pressure irrespective of water content; at 1173 K the difference in strength is ≈ 2 orders of magnitude for both wet and dry samples. At atmospheric pressure and high temperatures, *Dimanov et al.* [1999] observed fast dehydration of water present at grain boundaries. However, in the samples deformed at 300 MPa confining pressure, no significant dehydration was observed. In addition to the difference in water fugacity this difference in dehydration may also account for the different strengths of anorthite aggregates observed at 0.1 and 300 MPa confining pressure. *Xiao* [1999] also deformed synthetic wet anorthite

aggregates at 300 MPa confining pressure. His results are quite similar to ours (Table 3).

4.2.2. Activation energy. Farver and Yund [1995a] investigated Ca grain boundary diffusion in dry anorthite and oxygen grain boundary diffusion in wet albite. They found activation energies of 291 and 68-83 kJ mol⁻¹, respectively. Both values are substantially lower than the activation energies measured in this study, suggesting that neither the diffusion of calcium nor oxygen control grain boundary diffusion creep of dry and wet anorthite aggregates.

Recently, Yund and Farver [1999] reported activation energy of $176 \pm 61 \text{ kJ mol}^{-1}$ for Si grain boundary diffusion in anorthite aggregates. The diffusion experiments were performed at 0.1 MPa pressure, and the samples were considered dry from sample preparation, but no FTIR data are available. Al grain boundary diffusivities in feldspars are not known. Dresen et al. [1995, 1996] obtained activation energies of 365 and 425 kJ mol⁻¹ for grain growth of as-received (200 ppm H/Si) and dry (60 ppm H/Si) anorthite, respectively, at atmospheric pressure. These values are in the range of activation energies measured for diffusion creep of dry anorthite, but a comparison may not be straightforward since grain growth involves diffusion across the boundary as opposed to diffusion creep, which requires transport along the boundary.

4.3. The Effect of Water and Confining Pressure

A major difficulty in investigating plastic flow of silicates in deformation experiments at elevated pressures is to establish chemical equilibrium with respect to water [*Paterson and Luan*, 1990]. This requires that the kinetics of volume diffusion of water are fast enough to attain local equilibrium at experimental temperatures and that the chemical environment is well controlled. *Kronenberg et al.* [1986] investigated the mobility of



Figure 8. (a-c) Variation of log strain rate, Q, and log A with log water content of synthetic anorthite aggregates. Flow law parameters are given in Table 3. Strain rates were normalized to T = 1273 K, $\sigma = 100$ MPa, and d = 10 µm. They increase considerably with increasing water content. However, the activation energy Q and the preexponential constant (log) A strongly decrease with increasing water content. The given error bars are those of the flow laws, reflecting the statistical standard deviation of regression fits. (d) Plot of Q versus log A showing a linear interdependence and suggests a compensation law for diffusion (open symbols) and dislocation creep (solid symbols). The lines are calculated by linear regression of the creep parameters for anorthite obtained in this study. The results for other plagioclase rocks and for quartz and olivine are in reasonable agreement, suggesting that the SiO₄ tetrahedral structure with the incorporation of water determines the parameters of power law creep. See text for discussion.

hydrogen in potassium feldspar, and *Dimanov et al.* [1999] estimated diffusivities of molecular water in the same anorthite material as used in this study. Their results suggest that the diffusion distance at 1423 K is ≈ 2 mm for hydrogen and $\approx 4 \,\mu\text{m}$ for molecular water in 2 hours. For the range of temperatures of this study, even the smaller diffusion distance for molecular water is of the order of the average grain size of the anorthite aggregates.

Figure 8a is a plot of log strain rate-log water content. The data are from this study, *Xiao* [1999], and the deformation experiments at 0.1 MPa of *Dimanov et al.* [1999]. Dislocation

and diffusion creep strain rates in anorthite increase with water content at 300 and 0.1 MPa confining pressure, at least up to a water concentration of 0.1 and 0.05 wt %, respectively. A comparison of the data sets indicates that the effect of confining pressure appears to be small relative to changes in water content.

Paterson [1989] suggested that the dislocation creep rate increases with water fugacity $f_{\rm H2O}$. Provided that the rock is water saturated and in equilibrium with respect to the chemical potential of water, this may be expressed by a relation of the form

$$\dot{\epsilon} \approx f^p_{\ \text{H2O}}.$$
 (3)

For quartzite, Kohlstedt et al. [1995] and Gleason and Tullis [1995] inferred an exponent $p \approx 1.0$, suggesting that (2OH)₀ is the dominant point defect and an oxygen atom is replaced by a (OH)₂ group. Post et al. [1996] obtained $p \approx 2.0$ from their data, indicating that $(4H)_{S_1}$ is the major defect where four hydrogen atoms replace a silicon atom. Post et al. also demonstrated that quartz samples show a decrease in strength with increasing amount of water up to 0.3 wt %, with no further weakening for higher water contents. McDonnell [1998] recently found that fine-grained synthetic forsterite aggregates deformed triaxially with up to 0.1 wt % water present are substantially weaker than dry samples. On the basis of experimental data for quartz and olivine, Ord and Hobbs [1989] suggested that the activity exponent is less than the stress exponent since large changes in water activity lead to relatively small changes in flow stress at a given strain rate. Solution-precipitation processes appear to be relevant for creep under laboratory conditions only if more than ≈0.5 wt % water is present in fine-grained feldspar aggregates [Tullis et al., 1996; Tullis and Yund, 1991] and quartzite [den Brok et al., 1994; den Brok and Spiers, 1991].

The water fugacity in samples deformed in this study can be estimated from the water content measured using FTIR and the porosity of the samples. *Behrens* [1994] estimated the solubility of water at 300 MPa $P_{\rm H2O}$ and 1273 K to be \approx 7500 ppm H/Si. The water content in the dry samples is substantially less, but cooling stage FTIR measurements suggest that some of the inclusions contain freezable water. We consider it unlikely that the water precipitated during cooling and subsequent pressure release. However, *Doukhan and Trepied* [1985] suggested that the criterion of chemical equilibrium for isolated water bubbles is different from that for a constant pressure reservoir.

To estimate water fugacities, we assume that the total amount of water measured by FTIR is distributed in the pores. The ratio of pore volume to water content yields the specific volume of water for which the temperature dependent fugacity can be retrieved from steam tables [Burnham et al., 1969; Haar et al., 1984]. For the range of experimental temperatures and at a pressure of 300 MPa this procedure obviously yields only a rough estimate of the fugacities; they range from ≈ 10 MPa for dry samples to ≈ 150 MPa for wet specimens. Since we used an iron jacket, the oxygen fugacity is probably buffered by iron wustite. Using the modified Redlich-Kwong equation of state for H₂-H₂O fluid mixtures [Grevel and Chatterjee, 1992] predicts that the water fugacity at experimental conditions is $\approx 40\%$ lower than the first estimate.

For anorthite, water can be incorporated by the reaction $2(CaAl_2Si_2O_8) + 2xH_2O \leftrightarrow Ca_2Al_4Si_{4x}H_{4x}O_{16} + xSiO_2$, with x = 4 for the substitution Si⁴⁺ \leftrightarrow 4H⁺, or x < 4 for partial substitution. Alternatively, replacing O²⁻ by 2(OH)⁻ gives 2(CaAl_2Si_2O_8) + 2xH_2O \leftrightarrow Ca_2Al_4Si_4O_{16-4x}(OH)_{4x} + 2xO, again with x = 4 for complete replacement, or lower otherwise. In both cases, the fugacity exponent *p* equals *x*, suggesting *p* values between 0 and 4.

To interpret the effect of trace amounts of water and pressure on the creep rate of anorthite aggregates, at least two alternative approaches are conceivable. First, it may be possible that the activation energy for creep itself depends on water fugacity since the concentration of extrinsic water-related defects in thermodynamic equilibrium changes with fugacity. Second, the rate-limiting creep mechanism may change as a function of water content. A combination of both processes is possible.

Our observation is that the activation energies for diffusion and dislocation creep change with water content at least up to ≈ 0.1 wt % (Figure 8b). The data from experiments performed at 300 MPa confining pressure show reduced activation energies when compared to experiments performed by *Dimanov et al.* [1999] on samples with comparable water content deformed at 0.1 MPa pressure. However, at atmospheric pressure the water fugacity is poorly defined. For this reason we estimated the fugacity for our samples with different water content but a constant confining pressure of 300 MPa using a multilinear regression to the data. This procedure yields an estimated fugacity coefficient of ≈ 0.5 for both diffusion and dislocation creep. However, as already mentioned above, this approach is only valid if samples with low and high trace amounts of water deform with a similar rate-limiting mechanism, which has not been confirmed yet.

4.4. Isokinetic Effect

Figures 8b and 8c show that activation energy Q and log A decrease with increasing water content. Q and log A are linearly related for diffusion and dislocation creep (Figure 8d). For anorthite the Q-log A pairs obtained at different confining pressure and water content for a given creep regime form a straight line. The data from other studies on synthetic aggregates of feldspar (e.g., points 18 and 19 in Figure 8d), olivine (points 3, 5, 6), and quartz (point 4) fit to a similar trend provided that the stress exponents n are close to 3 and 1 for dislocation and diffusion creep, respectively. The data for quartz and olivine deviate somewhat from the regression lines determined for the anorthite samples, but they contain estimates of Q only (see Table 3). For olivine, data points 5 and 6 represent two different diffusion regimes with m = 2 and 3, respectively. However, *Hirth and Kohlstedt* [1995a] found m = 3 for olivine.

This kind of relation between Q and $\log A$ is known as the "compensation law" or "isokinetic effect" of diffusion. The temperature dependence of the diffusion coefficient D is usually expressed in the form of an Arrhenius law

$$D = D_0 \exp\left(-H/RT\right),\tag{4}$$

where D_0 is preexponential factor and H is activation enthalpy. The compensation law describes the linear relation

$$\log D_0 = \log D^* + H/2.303RT^*$$
(5)

where T^* and D^* are the "isotemperature" and "isodiffusion coefficient", respectively. T^* , D^* is the crossover point in a log D-1/T plot where the Arrhenius lines for different diffusing species in a given mineral, or for a given species in different materials, intersect [e.g., *Hart*, 1981; *Sorokin*, 1992]. For example, the compensation law may be applied to self-diffusion of ionic species in feldspar and olivine [*Hart*, 1981], silicon self-diffusion in silicates [*Bejina and Jaoul*, 1997], and divalent cation diffusion in olivine and garnet [*Jaoul and Sautter*, 1999].

The compensation law is still not well understood. It may be due to the common relationship between activation enthalpy Hand the preexponential factor D_0 to the Debye temperature, although it predicts a nonlinear interdependence [Lasaga, 1981]. Dowty [1980] and Zheng and Fu [1998] found a strong linear correlation between anion porosity and oxygen diffusivity for a wide range of minerals. Fortier and Giletti [1989] established a similar relation between oxygen diffusion in silicates and the total ionic porosity, suggesting that D_0 and H are both affected by the ionic porosity. Hart [1981] and Bejina and Jaoul [1997] explained the interdependence of H and D_0 by the relation between D_0 and the activation entropy S (of defect formation and migration). The compensation law then infers a relationship between S and activation enthalpy H. Bejina and Jaoul [1997] pointed out that D^* has a pure entropic significance, e.g., it is the value of D at H = 0.

A similar compensation behavior may be expected for diffusion and dislocation creep of feldspar since both creep mechanism are probably rate limited by Si or Al diffusion, with D_0 incorporated in the preexponential factor A. Performing a linear regression on our data points yields $Q = 281 + 28.9 \log A$ for dislocation creep of anorthite and $Q = 121 + 28.6 \log A$ for diffusion creep. From the slope b of these regression lines the compensation temperatures T^* are calculated from the relation T^* = b /2.303 R. T^* is ≈ 1510 K and 1495 K for dislocation and diffusion creep, respectively. Subdividing our data for diffusion of wet anorthite into temperature regions above and below 1343 K (see Figure 4a), results in parameters Q and $\log A$ that are in good agreement with the compensation line (points 9 and 10 in Figure 8d). Moreover, creep parameters of dry and wet anorthite match a single compensation line. We conclude that even when extrinsic (H₂O) defects change the activation energy for diffusion, the creep rate is probably still controlled by a single (rate limiting) species, i.e., Si or Al.

Bejina and Jaoul [1997] found a compensation law behavior for silicon self-diffusion in silicates as a first-order approximation, suggesting a common mechanism for Si migration in minerals with structures based on the SiO₄ tetrahedron. However, in detail, forsterite and quartz showed slightly different compensation lines, possibly related to their different crystallographic structures. A similar explanation may hold for the compensation behavior shown in Figure 8d. A difference in the structure between different silicate minerals may be responsible for the small deviation of quartz and olivine from the plagioclase compensation line.

5. Geological Applications

Feldspar is a major mineral constituent of the Earth's crust and often contains trace amounts of water [Beran, 1987; Hofmeister and Rossman, 1985]. Crustal shear zones commonly show evidence for deformation in the presence of aqueous fluids [e.g., Kenkmann, 1997; Kronenberg et al., 1990; Nakashima et al., 1995]. To extrapolate our experimental data to natural conditions, we focus on the deformation behavior of wet anorthite.

Passchier and Trouw [1996], Pryer [1993], and Tullis [1983] reviewed the deformation mechanisms operating in naturally deformed feldspars. A transition between dislocation creep accommodated by grain boundary migration recrystallization (low T) and by subgrain rotation recrystallization (high T) was suggested to occur under amphibolite facies conditions. In quartzo-feldspathic ultramylonites deformed at greenschist- to granulite-facies conditions, high ductility suggesting superplastic flow and diffusion creep accommodated by grain boundary sliding processes has been commonly observed [Allison et al., 1979; Jensen, 1985; Olsen and Kohlstedt, 1985; White, 1990; White and Mawer, 1986; Fliervoet et al., 1997; Kenkmann, 1997].

Figure 9a shows calculated flow stress profiles for wet anorthite in comparison with frictional strength determined by Byerlee's [1978] law. The friction stress for a thrust fault is given by $\sigma_{TF} = [2 \tau_0 \cos \phi + 2 S_v (1-\lambda) \sin \phi] / (1-\sin \phi)$, and for a normal fault by $\sigma_{\rm NF} = [2 \tau_0 \cos\phi + 2 S_v (1-\lambda) \sin\phi] / (1+\sin\phi)$, where τ_0 is cohesion (50 MPa), ϕ is angle of internal friction (31°), and λ is

Figure 9. (a) Extrapolation of wet anorthite dislocation creep data to natural strain rates of 10⁻¹³-10⁻¹⁵ s⁻¹. Also the intersection with friction strength calculated with Byerlee's friction law under hydrostatic pore pressure conditions is indicated. The mean crustal density was set to 2.8 g cm⁻³. TF is thrust fault (compressive) regime, NF is normal fault (extension) regime. Geothermal gradients of 20 and 30 K km⁻¹ are assumed. (b) Stress-temperature diagram of dislocation creep at 10⁻¹⁴ s⁻¹ strain rate for wet synthetic quartz, wet synthetic anorthite, and wet synthetic olivine. Power law parameters are given in Table 3.

ratio of pore pressure to overburden pressure. The transition from brittle frictional sliding to crystal-plastic flow of wet anorthite aggregates occurs at temperatures of ≈725-780 K for a strain rate of 10⁻¹⁴ s⁻¹, geothermal gradients ranging from 20 to 30 K km⁻¹, and a hydrostatic fluid pressure gradient (Figure 9a). The temperature at the transition changes by ≈25 K for ±1 order of magnitude change in strain rate. This temperature range is in good agreement with temperature estimates of 723-773 K for the onset of dislocation creep in naturally deformed feldspars based on microstructural observations [Voll, 1976; Pryer, 1993].

De Bresser et al. [1998] proposed that grain growth during diffusion creep and grain size reduction due to dynamic recrystallization accompanying dislocation creep may reach a dynamic equilibrium at the boundary of the two creep regimes. Extrapolated flow laws for dislocation and diffusion creep of wet anorthite intersect at a grain size of ≈15-70 µm for strain rates of

TF, 30 K/km NF, 20 K/kn NF, 30 K/kn 0 700 750 800 850 T, K stress. MPa 0 20 40 60 80 100 $\dot{\epsilon} = 10^{-14} s^{-1}$



 10^{-11} - 10^{-13} s⁻¹. That is in agreement with grain size distributions found for mylonites and ultramylonites in shear zones.

A comparison of strength profiles for quartz, feldspar, and olivine extrapolated to a strain rate of 10^{-14} s⁻¹ is given in Figure 9b. The effect of water fugacity on dislocation and diffusion creep rate of these minerals is not always well constrained. Therefore it was not considered in extrapolating the data. The flow laws are based on experiments performed on wet, fine-grained synthetic aggregates using a gas deformation apparatus at 300 MPa confining pressure. Considering these flow laws as the most reliable ones with respect to equilibrium with water, quartz is much weaker than plagioclase in the dislocation creep regime up to ≈ 1100 K temperature, and olivine is several times stronger even at 1200 K.

6. Conclusions

We examined the deformation behavior of fine-grained synthetic anorthite aggregates at 300 MPa confining pressure. Wet samples contain ≈ 0.07 wt % water, and dry specimens contain ≈ 0.004 wt % H₂O, incorporated as fluid inclusions and probably as point defects. Depending on the applied stress, temperature, and water content of samples, we observed a transition from diffusion creep with a stress exponent $n \approx 1$ to dislocation creep with $n \approx 3$. Microstructural observations of samples deformed in the dislocation creep regime suggest that grain boundary migration recrystallization is an important accommodation process. Grain boundary diffusion is expected in the low-*n* regime based on comparison with published creep experiments on similar material.

In the temperature range investigated dry samples are stronger than wet samples. The activation energy for both diffusion and dislocation creep of wet specimens is about 300 kJ mol⁻¹ lower than that for dry samples. This difference may be due to a gradual change from intrinsic to extrinsic water-related point defects controlling the creep rate or due to a dependence of Q on external variables (water fugacity) for these aggregates when only trace amounts of water are incorporated.

In the diffusion creep regime, anorthite deformed at 300 MPa pressure is considerably weaker than when deformed at 0.1 MPa, probably due to differences in water fugacity and due to different amounts of water species on grain boundaries. For dislocation creep we found higher activation energies in comparison to published data under dry and wet conditions, predicting higher strength at low temperatures.

A compensation law-like behavior was observed for dislocation and diffusion creep of plagioclase, similar to the isokinetic effect found in many diffusion experiments. This suggests that diffusion is controlled by the same ionic species (Si and/or Al) over a wide range of conditions in both regimes.

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