THE RHEOLOGICAL PROPERTIES OF ICE — WHERE WE ARE AND WHITHER WE ARE TENDING. D. L. Goldsby, Department of Geological Sciences, Brown University, 324 Brook Street, Providence, Rhode Island, 02912, David_Goldsby@brown.edu

Introduction: The dynamical properties of glaciers, ice sheets, polar ice caps, and icy planetary interiors are controlled in large part by the grain-scale deformation of ice. Here I review our current state of knowledge of the rheological properties of ice I, and identify crucial areas of current and future research that will provide fundamental constitutive laws and physical properties data required for modeling the dynamics of icy satellites and other cryospheric bodies.

Background: The rheological behavior of ice has classically been described by the Glen flow law, a power law relationship between strain rate and stress [1].

$$\dot{\varepsilon} = B\ \sigma^n \tag{1}$$

Here $\dot{\varepsilon}$ is strain rate, B is a constant at a given temperature, σ is stress, and n is the stress exponent (=3). The Glen law is predicated upon a large number of creep experiments on relatively coarse-grained ice (grain size > 1 mm) at relatively high temperatures (>250 K). Glen law flow is attributed to dislocation creep and deemed to be independent of grain size [2].

The flow of materials is often described by a generalized constitutive law of the form

$$\dot{\varepsilon} = A\ \sigma^n\ \frac{1}{d^p} \exp\left(-\frac{Q_c}{RT}\right) \tag{2}$$

where A is a materials parameter, d is grain size, p is the grain size exponent, Q_c is the activation energy for creep, R is the gas constant and T is temperature. Major advances in the laboratory within the past two decades [3-9] have demonstrated unequivocally that the rheological behavior of ice is much more complex than suggested by the simple Glen law. Ice flows via any of four creep mechanisms, each characterized by a flow law of the form of Eq. 2, and each of which limits the overall creep rate over different conditions of σ, T, and d. These studies have demonstrated, in fact, that the Glen law represents transitional behavior between dislocation creep (characterized by n=4) and grain boundary sliding-accommodated dislocation creep (characterized by n=1.8) [3], and thus cannot be meaningfully extrapolated to stresses outside the limited range explored in the laboratory experiments considered by Glen.

Dislocation creep - An exhaustive study of the flow of ice at elevated confining pressures, conducted primarily with planetary applications in mind, reveals that coarse-grained ice (d >250 μm) deforms via *dislocation creep*, characterized by $n \approx 4$, over a wide range of σ and T [7-9]. This $n \approx 4$ behavior is also observed in 1-atm experiments on fine-grained samples (d <100 μm) at stresses >1 MPa [3]. Fine grain sizes prevent microcracking at these stresses at ambient pressures. Excellent agreement between creep data from high-pressure experiments on coarse-grained samples and from ambient-pressure tests on fine-grained samples [3] demonstrates that dislocation creep of ice is independent of grain size (i.e., p=0). In accordance with the von Mises criterion for compatible deformation of polycrystalline solids, dislocation creep of ice cannot occur solely via dislocation slip on the easily activated basal slip system, and requires activation of more creep resistant, non-basal slip systems [10].

Grain size-sensitive (GSS) flow – The flow of materials at low stresses and high temperatures is controlled by GSS creep processes, wherein the creep rate increases with decreasing grain size [11]. Such GSS creep mechanisms control the creep rate in low differential stress natural environments, such as the interiors of the icy satellites. The grain size dependence embodied in Eq. 2 is a natural consequence of grain boundary sliding (GBS), whereby neighboring grains in a material are translated along their mutual grain boundaries. To provide compatible deformation of polycrystalline materials, GBS must be accommodated by dislocation motion or diffusional flow.

Mutually accommodating GBS and dislocation slip - Laboratory experiments on fine-grained ice samples (d = 3-200 μm) [3-5] reveal that over a broad range of conditions compatible flow of ice occurs via slip on the basal dislocation slip system acting in concert with GBS, with the overall creep rate limited by the slower of these two processes. For *GBS-accommodated basal slip creep*, characterized by n=1.8 and p=1.4, the creep rate is limited by GBS, whereas for *basal-slip accommodated GBS creep*, characterized by n=2.4 and p=0, the overall creep rate is limited by basal slip [3]. Extrapolations of the flow laws for each of these creep mechanisms to the larger grain sizes characteristic of natural ice bodies demonstrates that GBS-accommodated basal slip creep, hereafter termed GBS creep or GBS flow for brevity, rate limits the flow of ice over a broad range of conditions overlapping those in glaciers, ice sheets, and icy planetary interiors [3]. The GBS law has been employed in modeling a host of glaciological and icy planetary phenomena [12-16].

Outstanding Problems: In spite of the great advances in our understanding of the rheological properties of ice within the last two decades and the successful application of the GBS creep law in models of

various planetary phenomena, crucial gaps remain in our knowledge. Outstanding problems include 1) the nature and extent of the diffusion creep regime for ice, 2) the influence of impurities on ice flow, particularly for grain size-sensitive flow mechanisms, and 3) the relationship between ice flow and the development of lattice preferred orientation (LPO).

Diffusion creep - For a given grain size and temperature, diffusion creep should dominate the rheological behavior of ice at stresses lower than for GBS creep [17], and therefore should compete with that mechanism for dominance in low stress planetary environments. Recently, the role of diffusion creep as a rate-controlling creep mechanism in convective processes within icy satellites has been emphasized [e.g., 18]. In spite of its importance, however, the diffusion creep regime for ice has never been accessed in laboratory experiments [3]. Experimental validation of diffusion creep for ice and quantification of the diffusion creep rate would represent major advances in our understanding of planetary ice flow.

The diffusion creep rate is given by the classical diffusion creep equation [11]

$$\dot{\varepsilon} = \frac{C\,\Omega\,\sigma}{d^2 RT}\left[D_v + \frac{\delta D_b}{d}\right] \qquad (3)$$

where C is a constant which depends on grain geometry, Ω is molar volume, k is Boltzmann's constant, D_v is the volume diffusion coefficient, δ is the grain boundary width, and D_b is the grain boundary diffusion coefficient. The diffusion coefficients are of the form $D=D_o\, exp(-Q_d /RT)$, where D_o is a constant and Q_d is the activation energy for diffusion. Volume and grain boundary diffusion contribute independently to the creep rate, i.e., are parallel kinetic processes, yielding a $1/d^2$ dependence of the creep rate when volume diffusion dominates, and a $1/d^3$ dependence when boundary diffusion dominates. A hallmark of diffusion creep is the linear dependence of creep rate on stress.

Diffusion creep and GBS creep are parallel kinetic processes, such that the faster process dominates the creep rate. Comparison of diffusion creep rates calculated via Eq. 3, using known and estimated equation parameters, with GBS creep rates for all possible combinations of σ, T, and d indicate that in order for diffusion creep to be *faster* than GBS creep or basal-slip limited creep, samples with d <3 µm must be deformed at T >248 K, a practical impossibility due to rapid grain growth. The diffusion creep regime will therefore likely never be accessed in the laboratory for pure ice samples, underscoring the need for measurements of the grain boundary diffusion coefficient in ice, the sole unknown in the diffusion creep equation, as a means of constraining the diffusion creep rate.

Effects of impurities on ice flow – Despite the successful application of the GBS and dislocation creep laws in modeling icy planetary phenomena, important questions remain about the effects of particulates, likely common in icy satellites, on the rheological behavior of ice. The effects of particulates on dislocation creep of ice remain poorly constrained; the effects of intergranular and intragranular particulates on GBS flow of ice are unknown and are the subject of current research. Similarly, the effects of ionic impurities segregated to grain boundaries on ice flow have never been investigated for any ice creep mechanism. Grain growth kinetics for particulate-laden ices also require quantification, and are the subject of current research.

LPO development during GSS creep – Unlike the icy satellites, glaciers and ice sheets provide a natural laboratory where theories and predictions about ice flow in nature can be tested directly. Comparison of GBS flow laws with rheological data from glacier field experiments strongly suggests that under most conditions, the flow of these natural ice bodies occurs via GBS creep [3, 19]. Microstructures from samples deformed by GBS creep are also remarkably similar to those observed in glaciers and ice sheets [3, 19]. Despite the excellent agreement between mechanical and microstructural data for GBS creep from the laboratory and those from field studies, however, important questions remain regarding the development of LPO during GBS flow. Ice crystals in glaciers and ice sheets display strong LPO. Conventional wisdom holds that creep mechanisms involving GBS do not lead to LPO. However, when dislocation slip and GBS are mutually accommodating and significant strain occurs via dislocation slip, as in the GBS creep regime for ice, LPO development is expected theoretically, and observed for other materials [10]. LPO development during GBS creep of ice is a subject of active current research.

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