DOES THE BRITTLE-TO-DUCTILE (MOBILITY) TRANSITION OF ICY SAND PACKS COINCIDE WITH THE MAXIMUM PACKING DENSITY?

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Summary: We present preliminary experimental data indicating a correspondence between the brittle-to-ductile transition of icy sand packs and the maximum packing density of sand in such packs. If confirmed, such a relationship would have important ramifications for both terrestrial and extraterrestrial ice. Based on laboratory experiments performed at 263 K, Mangold et al. [1] concluded that the brittle-ductile transition (which we call here more descriptively the "mobility" transition) in ice-rock mixtures occurs at ice volume fractions lower than 0.28. However, we have previously demonstrated the importance of conducting ice deformation experiments at colder temperatures more relevant to planetary conditions [2], thereby providing the initial motivation for the current study.

Experimental Setup: Creep experiments on ice + sand mixtures were carried out at elevated pressure P and low temperature T in the same cryogenic gas creep apparatus used and described by [2]. We measured the rate of axial shortening of cylindrical samples under a state of constant differential stress σ, which in conservative adherence to the "Goetze rule" was not allowed to exceed half of the confining pressure. Samples are composed of sand, triple-distilled water ice, and in some cases air-filled porosity vented to room pressure. For the purposes of this work we characterize composition by the relative volumes of sand, f, and ice, $1 - f$, and specify the degree of ice saturation. The primary sand utilized is Oklahoma #1 (OK #1), a high-purity quartz sand. OK #1 has a narrow distribution of grain size, with 84% of the grains having a diameter between 0.106 and 0.25 mm. The loose porosity of OK #1 is $\phi = 0.42 \pm 0.02$; with shaking, it reduces to as low as $\phi = 0.32 \pm 0.02$. For comparison, the theoretical minimum porosity for spheres of uniform grain size in a "maximally random jammed" state is 0.36 [3].

Results: We carried out an initial series of runs on samples of composition from $f = 0.50$ to 1.00, with the sand portion being purely OK #1. The primary independent variable was sample composition, in keeping with our focus on the mobility transition. Most of the experiments were carried out at 223 K under a hydrostatic pressure of 60 MPa and a differential stress near 30 MPa. All samples displayed an initial period of work hardening, wherein strain rate decreased with strain ε. For two samples with high ice content ($f \leq 0.60$), the work hardening stage was followed by apparent steady-state ductile deformation (consistent with our earlier ice-rich results [2]), wherein strain rate no longer evolved with strain.

Conversely, samples with lower ice content ($f \geq 0.8$), which were necessarily undersaturated in ice, work hardened without limit; that is, strain rate decreased monotonically with strain and eventually dropped below our limit of observation, approximately 3×10^{-9}/s. We associate such behavior with the brittle field of deformation, since further deformation cannot proceed by plastic flow alone, but requires either cataclasis or dilatancy. In these ice-poor samples, the most important factor in work hardening seems to be the presence of ice in any amount. The actual concentration of ice has little effect on the shape of strain-time curves; however, removing the ice brings strain rate to a halt in a much shorter time. Evidently, even a small amount of water ice facilitates a rearrangement of sand particles under differential load.

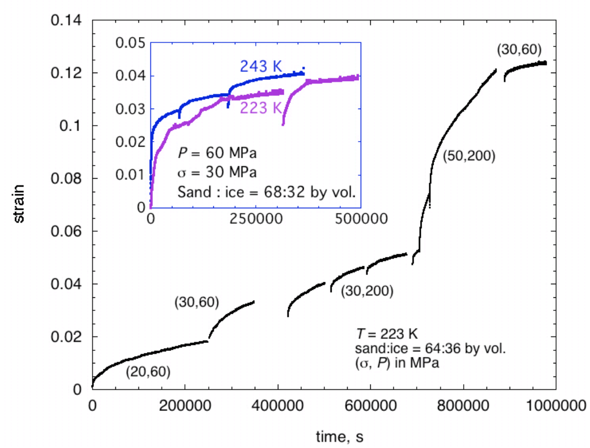

Figure 1. Transitional run, $f = 0.64$, saturated, with its five run steps labeled by differential stress and pressure in MPa. Note that the third and fourth steps at $P = 200$ MPa are nominally in the stability field of ice II. Inset: contrasting behavior of two samples with only slightly higher sand content, $f = 0.68$, causing them to fall just short of full mobility.

Transitional runs: Fig. 1 shows the evolution of strain rate near the mobility transition. In steps (1) and (2) of the saturated $f = 0.64$ run, the sample exhibits the same constant hardening seen in sand packs that are not fully mobile, although at a strain of 0.03 near the end of step (2) the strain rate (corresponding to the slope of the curve) is much higher than that of samples of $f = 0.68$ at the same conditions (see inset). Steps (3) and (4) do not show a loss of ductility and in fact deform far beyond the strain of ~0.04 that seems to limit the other sand packs. Step (5), back at the "common" conditions of σ = 30 MPa and P = 60 MPa, seems to confirm that the *f* = 0.64 sand pack is indeed mobile.

Now consider the two ice-saturated *f* = 0.68 runs (Fig. 1, inset) at *T* = 223 K (purple) and *T* = 243 K (blue; also shown in Fig. 2): the limiting strain is roughly 0.04, about the same as for the ice-poor runs (*f* ≥ 0.8), but samples take far more time to reach that limit (Fig. 2). Also, changing temperature or stress at fixed *f* appears to change the rate at which the limiting strain is approached, but does not change the limiting strain itself. For example, increasing *T* from 223 K to 243 K in a ductile sand pack (*f* = 0.50 and 0.60) changes the rate of strain by a factor of more than ten [2]; the same temperature change in the *f* = 0.68 ice samples has some effect on strain rate at early time, but virtually no effect on the total strain imparted (Fig. 1, inset).

Consequently, we conclude that the mobility transition at *T* = 223 K must occur between the brittle *f* = 0.68 and the ductile *f* = 0.64. CryoSEM images of the *f* = 0.64 sample show that while some sand grains are fragmented, the vast majority are intact, indicating that the deformation of the sand pack was ductile and not an artifact of mechanisms that can only occur at very high pressures. Therefore, we infer that the mobility transition at *T* = 223 K occurs closer to *f* = 0.68 than *f* = 0.64, resulting in an estimated ice volume fraction of 0.33 ± 0.01 at the boundary.

Implications: Note the similarity between the maximum packing density of dry OK #1, which was a volume fraction of 1–ϕ = 0.68 ± 0.02, and the volume fraction of OK #1 at the mobility transition under saturated conditions, *f* = 0.67 ± 0.01. One possibility is that the sand fraction at the transition is absolute, and that the close correspondence to maximum dry packing density is coincidental. Alternatively, it may be that the mobility boundary has a causal relationship with dry packing density. To investigate this connection, we carried out additional tests on sand packs where higher dry packing densities could be achieved; namely, in mixed sands of dissimilar grain sizes.

Fig. 3 shows the deformation history at 243 K of a saturated *f* = 0.71 sand pack that is a 2:1 (by volume) mixture of Flint Silica #12 (1-2 mm grain size) and OK #1 (0.1- 0.25 mm). This combination of sands was chosen after numerous iterations of sand mixing and shaking; it has a maximum dry packing volume fraction of 1–ϕ = 0.75. As seen in Fig. 3, this sand pack is unmistakably mobile. In contrast, the saturated *f* = 0.68 sample (containing only OK #1 and ice) shown in Fig. 2 is clearly not mobile under identical *T* and *P* conditions. This is compelling evidence that the mobility transition does not occur at a specific sand packing density, but rather is related to maximum packing density characteristic of the particular sand.

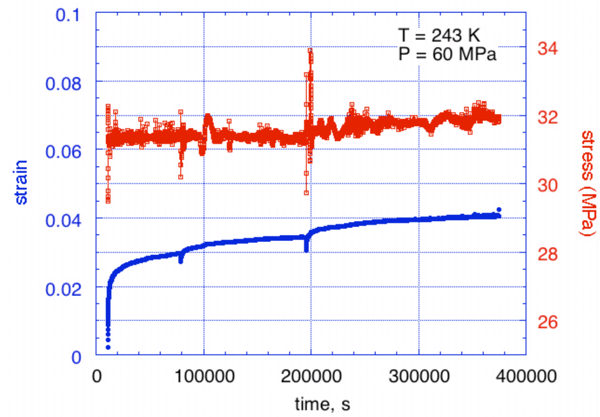

Figure 2. Saturated sand pack, *f* = 0.68. Note that strain rate (the slope of the blue curve) asymptotically approaches zero, indicating that the sand pack is immobile. The offsets in the blue curve are real strain enhancements that occur when the stress is removed and reapplied.

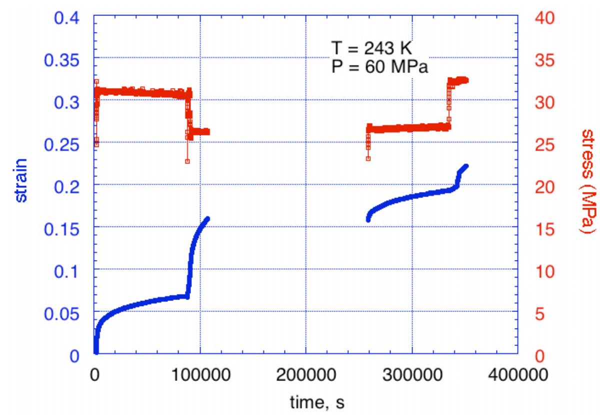

Figure 3. Mixed sand run, *f* = 0.71. High strains and the achievement of a steady-state strain rates indicate that this sand pack is highly mobile, even though the volume fraction of sand is higher than in Fig. 2. The difference here is a higher achievable dry packing density.

If confirmed by our further experiments on samples with high sand fractions, this result could have significant planetary implications. For example, the grain size distribution of the impact-pulverized Martian regolith may allow for mobility at much lower ice content than assumed by [1], which would dramatically decrease estimates of Mars' global volatile inventory.

References: [1] Mangold N. P. et al. (2002), *PSS*, *50*, 385-401. [2] Durham W.B. et al. (1992) *JGR*, *97*, 20883-20897. [3] Torquato, S. et al. (2000) *Phys. Rev. Lett.*, *84*, 2064-2067.