Bend gliding (Fig. 20) is presumably initiated after some elastic distortion, and the kink band is formed as the material buckles. According to Barrett (Ref. 94, p. 375), the kink band boundaries are initially planes that are normal to the gliding direction and the gliding plane. It follows, therefore, that well-developed kink band boundaries are initially oriented at high angles to the greatest principal stress.

Kink bands developed in biotite crystals of experimentally deformed Westerly granite(43) demonstrate their potential usefulness as a dynamic petrofabric element. Cylinders of the granite were compressed 15 to 23 per cent under 5-kb confining pressure at 500°C. Most of the shortening was achieved by faulting, localized along a zone of mylonitization. Biotite crystals near the shear zone were deformed by kink banding. The individual kink bands are described as being narrow and sharply bounded by nearly planar surfaces trending at high angles to the basal cleavage (001). They are conspicuous because the cleavage is sharply deflected through angles of 20 to 50 degrees. Opposite boundaries of a kink band typically are nonparallel, so that the band is wedge- or lens-shaped. In some grains several subparallel kink bands are present. In others there are two sets, symmetrically inclined to (001) of the host grain, giving a chevronlike pattern (Fig. 36(a)). Normals to the kink band boundaries tend to group around the known compression axis, σ_1 (Fig. 36(b)). Moreover, kink bands tend to develop preferentially in grains whose [001] axes (normals to (001)) are steeply inclined to σ_1 (Figs. 36(c) and 36(d)). That is, the (001) cleavages (the active gliding plane) are subparallel to σ_1. Subsequent work by Handin and Borg(53) on Fordham gneiss has substantiated these findings.

Accordingly, if biotite crystals in a naturally deformed rock are nearly randomly oriented, it is reasonable to conclude that the center of concentration of normals to kink band boundaries can be equated to σ_1 in the rock at the time of deformation. On the other hand, as biotite in most rocks exhibits a nonrandom orientation, this technique can position the greatest compression axis only somewhere within the s-plane defined by the basal cleavages.

Quartz Deformation Lamellae

General. Quartz deformation lamellae are here treated independently of any one mode of origin because the process by which they are formed in naturally deformed rocks is still somewhat in doubt. Sufficient information is available, however, to utilize these features in dynamic analyses. They occur in deformed quartz in a variety of rock types. Usually grains with deformation lamellae also show undulatory extinction. Their physical appearance is variable (Fig. 38 and Fig. 42 on p. 521). Most workers agree, however, that their tendency to lie at low angles to the $\{0001\}$ plane in the quartz host is diagnostic (Fig. 39).

Origin. The origin of quartz deformation lamellae has been the subject of much speculation since they were first reported 85 years ago. According to Griggs and Bell(42) and Ingerson and Tuttle,(130) Kalkowsky(131) first described the lamellae; Boehm(132) identified them as planes of liquid inclusions; Judd(133) thought they were secondary twin lamellae; and Becke(134) identified them as healed fractures. Mügge,(135) Sander,(1) Hietanen,(136) Fairbairn,(137,138)

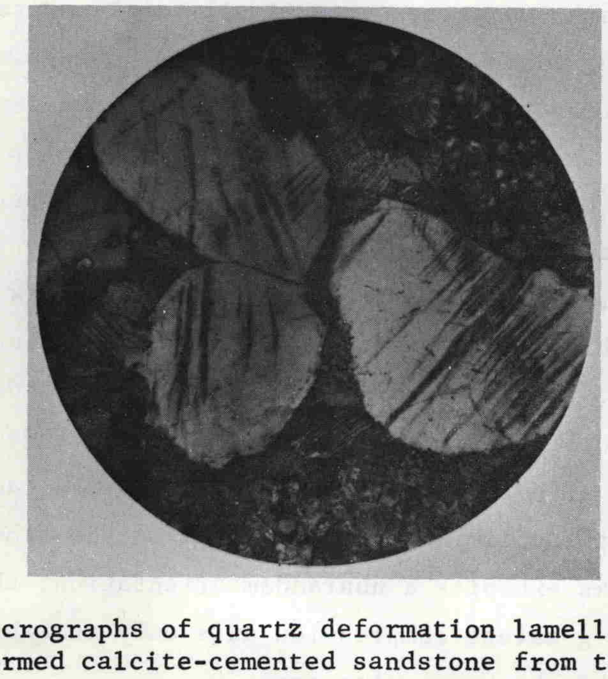

Fig. 38—Photomicrographs of quartz deformation lamellae in grains of a naturally deformed calcite-cemented sandstone from the Jurassic Piper formation, Park County, Montana. Note two sets of lamellae in one grain. Crossed Nicols, X 100.