SIR,

Some additional remarks on isotope stratification of the snow cover in high mountains

The deuterium and oxygen-18 contents in snow and ice of the upper layers of high mountain glaciers were studied earlier by us in two crevasses in the Peruvian Andes and in the Himalaya (Grabczak and others, 1983). As a result of this study, combined with simplified modelling, we concluded that the observed significant enrichment in both D and ^{18}O contents of the uppermost snow layer is due to its partial melting and/or sublimation caused by both the sun and winds. Comparison of the D and ^{18}O data suggested that these processes proceed under conditions close to the thermodynamic equilibrium; additional kinetic fractionation normally observed for evaporation from open water bodies seems to be negligible.

In support of our previous conclusions, we have undertaken an additional isotopic study of an upper layer of snow cover at two different altitudes on the north-eastern slopes of Aconcagua, the highest peak of both Americas (6959 m a.s.l., lat. 32.65°S., long. 70.00°W.). (The snow samples were collected by the members of the expedition to Aconcagua organized by the Krakow Branch of the Polish Society of Earth Sciences in January–February 1985, and were analysed for D and ^{18}O content in our laboratory.) These data should also contribute to a better understanding of the processes leading to isotope differentiation of the snow cover in a high-mountain environment, especially in view of recent attempts to extract paleoclimatic information from high-mountain glaciers of both hemispheres (Schotterer and others, 1978; Thompson and others, 1984, 1985).

The first group of profiles (four sets of samples) was taken on a small snowfield just below the glacier De los Polacos (altitude of *c.* 5200 m a.s.l.). The maximum depth of each profile of this group was about 60 cm with a sampling interval of 5–10 cm. Two profiles (Nos I and III) represent that part of the snowfield with exposure to the Sun, while the remaining two (Nos II and IV) were taken from a less-exposed area. The distance between individual profiles of this group was 4–6 m.

The second group of samples was collected about 50 m below the summit of Aconcagua, at an elevation of *c.* 6910 m. Two profiles, 90 cm in depth, were taken from the southern and northern slopes of a small mountain range, 10 m from each other (profile Nos V and VI, respectively).

The results for the ^{18}O content found in the analysed samples are summarized in Figure 1. Variations along profile Nos I to IV are quite similar: an initial sharp decrease in ^{18}O (and D) content is followed by a roughly constant value in the lower part. In two profiles (Nos I and III, both of which were more exposed to the Sun), a slight minimum at 20–30 cm can be distinguished. Profile No. I differs from the others possibly because of the absence of the uppermost layer, leading to a shift in the remaining data points in the profile. Relatively small differences between the profiles are remarkable below a depth of 40 cm, i.e. for the firn samples.

The fact that the surface snow in a high-mountain environment may differ in its ^{18}O content by even more than 20‰, when compared to the underlying strata, confirms our earlier suggestions (Grabczak and others, 1983) that the isotopic composition of the snow cover can be significantly modified by partial melting and sublimation processes. Water resulting from the melting of snow, and depleted in heavy isotopes by isotopic fractionation, percolates down to deeper layers and can be trapped by the glacier snow during its firnification. The degree of the isotopic enrichment of the surface-snow samples suggests that quite a significant proportion of the snow cover has been removed.

The profiles sampled near the summit of Aconcagua reveal completely different isotopic characteristics when compared to the other profiles. There are only very small fluctuations about the ^{18}O content (equal to -22.2 ± 0.2‰ and -24.2 ± 0.3‰ for profile Nos V and VI, respectively). The samples from sites with a more intense exposure to the Sun tend to be heavier in ^{18}O. The absence of isotope differentiation in the surface snow suggests that either the isotopically modified (by sublimation) surface snow layer has been removed by strong winds blowing on Aconcagua or that, for the same reason, the sublimation process takes place within the whole porous structure of the snow cover at the same rate. The significance of snow sublimation in the modification of the heavy-isotope content is evident when the ^{18}O data for the "summit" profiles are related to the results corresponding to the snow-pack samples from profile Nos I to IV. Specific meteorological conditions prevailing on the high mountain peaks may result in a completely different isotope composition of the snow when compared with the mountain slopes; this is known as the "summit effect".

The positions of the data points for the snow samples

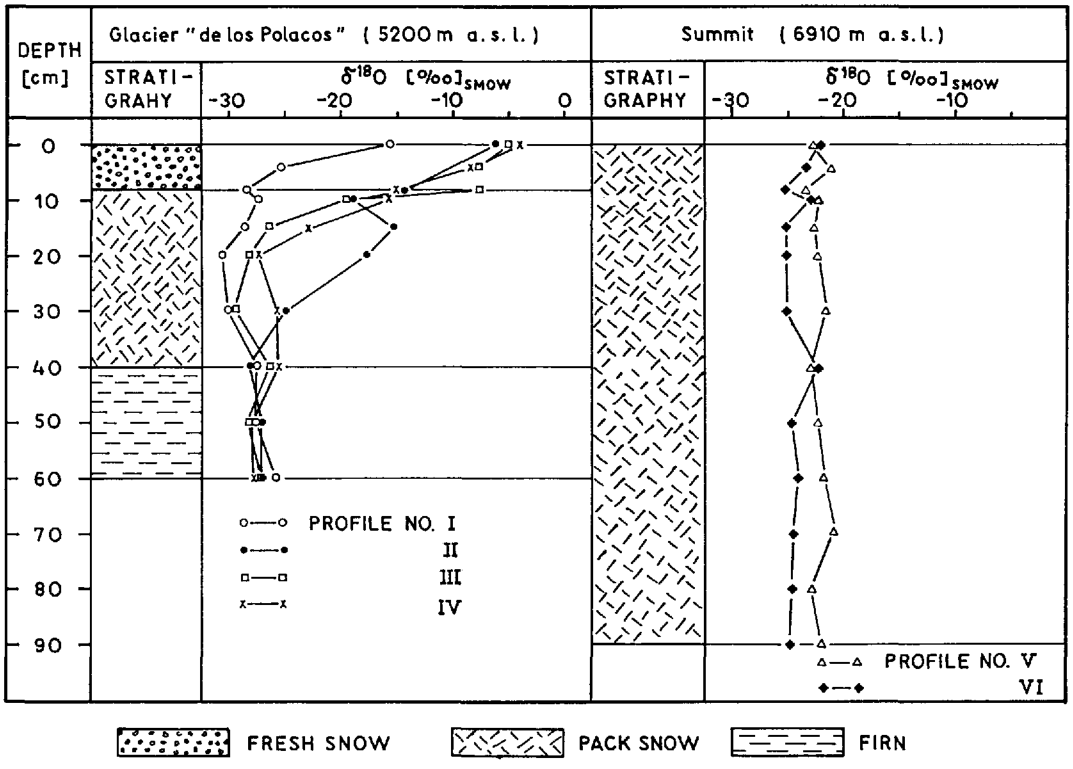

Fig. 1. ^{18}O content versus depth for a snow cover sampled on Aconcagua.

from all the sampled profiles on the δD versus $\delta^{18}O$ diagram follows the Meteoric Water Line with a slope close to 8, which argues for isotope equilibrium conditions during transformation from snow to firn.

The above results, together with our earlier findings and the data of other authors (Thompson and others, 1984), provide clear evidence that the snow cover in a high-mountain environment undergoes significant isotope modification prior to its tranformation into ice. The potential usefulness of high-mountain glaciers as a source of paleoclimatic information depends critically on whether isotope variations in precipitation induced by fluctuations of climate can somehow survive the firnification process.

Institute of Physics and Nuclear Techniques,
Academy of Mining and Metallurgy,
al. Mickiewicza 30,
30-059 Krakow, Poland

JERZY GRABCZAK
JERZY NIEWODNICZAŃSKI
KAZIMIERZ RÓŻAŃSKI

8 August 1988

REFERENCES

Grabczak, J., J. Niewodniczański, and K. Różański. 1983. Isotope stratification in high mountain glaciers: examples from the Peruvian Andes and Himalaya. *J. Glaciol.*, 29(103), 417-424.

Schotterer, U., W. Haeberli, W. Good, H. Oeschger, and H. Röthlisberger. 1978. Datierung von kaltem Firn und Eis in einem Bohrkern vom Colle Gnifetti, Monte Rosa. *Jahrb. Schweiz. Naturforsch. ges. Wiss. Teil*, 48-57.

Thompson, L.G., E. Mosley-Thompson, P.M. Grootes, M. Pourchet, and S. Hastenrath. 1984. Tropical glaciers: potential for ice core paleoclimatic reconstructions. *J. Geophys. Res.*, 89(D3), 4638-4646.

Thompson, L.G., E. Mosley-Thompson, J.F. Bolzan, and B.R. Koci. 1985. A 1500-year record of tropical precipitation in ice cores from the Quelccaya ice cap, Peru. *Science*, 229(4717), 971-973.

ERRATUM

Vol. 33, No. 115, p. 285, Fig. 5

The incorrect illustration for Figure 5 was inadvertently included in this paper. The following is the correct illustration.

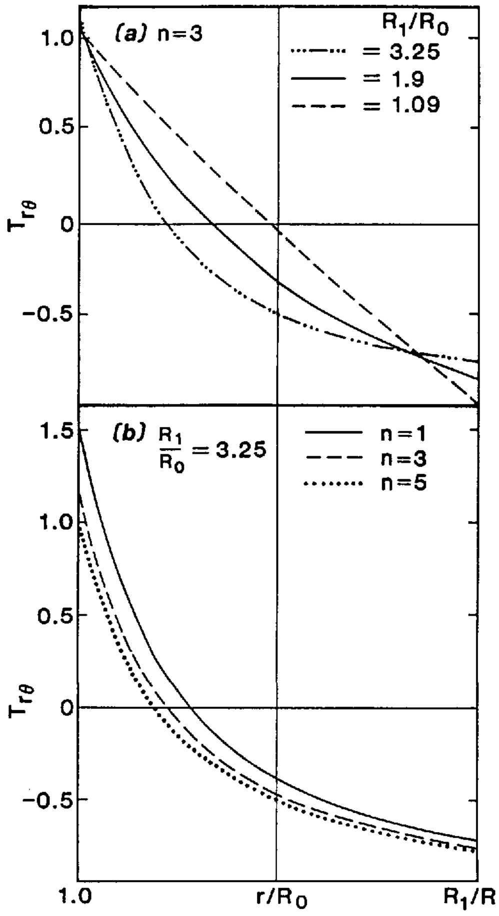

Fig. 5. Normalized shear stress ($T_{r\theta}$) across the width of a curving, rectangular channel with zero traction at the bed, for: (a) different values of the curvature, and (b) different stress exponents. Curves are obtained from the analytical discussion in the text.