Fig. 3. General view towards the south of the relict glacier of Jou Negro (Macizo Central). The highest peak of the massif (Torre de Cerredo, 2651 m), outside the photograph, lies just to the left of the margin (taken on 20 September 1992).

comparable to that described by Serrat (1980) in Madaleta Massif (central Pyrenees), where important glacier retreat took place from the 19th century until 1957. According to Serrat, the best-developed moraines, formed in the 19th century, are separated from the present glacier margins.

Besides Glaciar Jou Negro, it is possible that there is glacial ice beneath the northeast snow patch of Torre de la Palanca and the snow patch of Llambrión, both in the Macizo Central. Some snow patches in the Macizo Occidental could also have ice cores, such as Neverón de la Forcadona, on the north face of Peña Santa de Castilla, although the altitudes of the summits here are slightly lower than in the Macizo Central.

It is not surprising that glacial ice has not been seen until now in Picos de Europa as it is usually beneath a layer of snow and firn that acts as an insulator. Due to the low precipitation during the last few years, ablation exceeded precipitation, causing melting of most of the snow and firn. As a result, the ice cores have been exposed in recent summers. In September 1992, Glaciar Jou Negro was discovered by one of the authors (González Suárez).

The preservation of these small glaciers in Picos de Europa can be attributed to heavy precipitation, in the form of snowfall and the permanence of this snow favoured by low solar radiation. As perennial snow-cover provides insulation, ablation was considerably reduced, and this allowed the continued existence of the glacial ice. The present state of these glaciers is not known, although future surveys will enable us to establish whether they are stable or whether they are retreating.

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SIR,

Comments on "Subglacial floods and the origin of low-relief ice-sheet lobes" by E. M. Shoemaker

Shoemaker (1992) proposed that gently sloping lobes of the Laurentide ice sheet (e.g. Mathews, 1974; Beget, 1986) were not steady-state features due to ice movement over weak, deformable sediment, but rather were transient features whose advance was triggered by giant subglacial floods of water, moving in a meters-thick sheet, released from subglacial storage. Although there is much geological evidence for episodic advance of gently sloping ice-sheet lobes (e.g. Clayton and others, 1985; Clark, 1993), Shoemaker's proposed explanation is untenable, because thick water sheets are unconditionally unstable to formation of channels (Walder, 1982). In the remainder of this commentary, I elaborate this criticism and also remark on other problems in Shoemaker's discussion of subglacial hydrology.

Nye (1976, p. 207) briefly considered whether an outburst flood might take the form of a water sheet. He concluded this was unlikely for two reasons. First, unless both the ice surface and the bed have no lateral slope whatsoever, water flowing in a sheet tends to be driven into channels. Secondly, lateral variations in water-sheet thickness tend to be accentuated by concomitant variations in frictional melting of the basal ice. Nye's remarks on the latter issue motivated my analysis (Walder, 1982), demonstrating that sheet flow is unconditionally unstable to formation of channels. I went on to present a heuristic argument that the effect of bed roughness might nevertheless allow sheets up to a few millimeters in thickness to be quasi-stable. Weertman and Birchfield (1983) subsequently argued that channelized flow itself is unstable if all meltwater is subglacially derived, due to the supposed inability of channels to collect water from large lateral distances. I am skeptical of this conclusion for several reasons; for present purposes, it is sufficient to note that their result applies only for rigid, impermeable beds. If the bed consisted of permeable sediment, as was certainly the case for marginal lobes of the Laurentide ice sheet, there would be no impediment to lateral water movement to channels. Thus the meters-thick water sheets proposed by Shoemaker are undoubtedly unstable to channel formation. Oddly, Shoemaker (1992, p. 107) cited my 1982 analysis but failed to note its fundamental conclusion.

A second instability leading to channelization is likely to arise as well for parts of an ice sheet flowing over unconsolidated sediment, such as the marginal lobes of the Laurentide ice sheet. This instability would be formation of channels *cut into the sediment*. Subaerial sheet wash over a hillslope is unstable to rill formation (Smith and Bretherton, 1972; Loewenherz, 1991), and I expect a similar phenomenon to occur subglacially (see also Walder and Fowler, 1994).

Shoemaker (1992, p. 107) asserted that channelized subglacial water flow is unstable relative to sheet flow if the ice-surface slope decreases in the downstream direction. He stated that "the reduction in the pressure gradient renders the ... channel ... incapable of carrying the discharge ... a water sheet is created". This statement seems to be based on the implicit, but erroneous, assumption that channel dimensions are fixed. The steady-state relation between discharge Q, cross-sectional area S and water pressure p_c may be written, following Nye (1976, p. 189), as

$$\rho_w g \sin\theta - \frac{\partial p_c}{\partial s} = \frac{NQ^2}{S^{\frac{8}{3}}} \tag{1}$$

where ρ_w is the density of water, g is acceleration due to gravity, θ is the bed slope, s is a coordinate along the channel and N depends on channel roughness and shape (but not size). The expression on the lefthand side of Equation (1) is the total hydraulic gradient. It is well known (e.g. Shreve, 1972; Weertman, 1972; Walder and Fowler, 1994) that, in the case of gently sloping ice and bed, the hydraulic gradient is dominated by the ice-surface slope:

$$\rho_w g \sin\theta - \frac{\partial p_c}{\partial s} \approx \rho_i g \sin\alpha \tag{2}$$

where ρ_i is the density of ice and α is the ice-surface slope. Equation (2) does not apply near the terminus. Substituting Equation (2) into Equation (1) gives

$$\rho_i g \sin\alpha \approx \frac{NQ^2}{S^{\frac{8}{3}}}\,. \tag{3}$$

Thus, a downstream decrease in α can be accommodated by an increase in S. More generally, when the approximation (2) does not hold, the pressure gradient as well as the channel area may change in a reach where $\partial\alpha/\partial s < 0$. Shoemaker's (1992) conclusion — that water must leak out from a channel if the ice-surface slope decreases downstream — is incorrect.

Finally, I wish to note Shoemaker's (1992) improper application of Nye's (1976) approximate analysis of outburst-flood hydrographs. Nye elegantly showed that the rising limb of at least some outburst-flood hydrographs could be calculated by neglecting plastic closure of the outlet tunnel. To predict the entire hydrograph, it is necessary to consider plastic closure and other factors, including temperature of released water and the shape of the water reservoir (Clarke, 1982). Shoemaker (1992, p. 112), however, ignored plastic closure and other possible effects for the entire duration of the putative outburst. He used values of discharge so calculated (along with the erroneous conclusion that flood waters leak out of the tunnel to form a sheet) to determine the period of time during which flood water supposedly would form a meters-thick sheet at the glacier bed and thereby trigger rapid ice advance.

In conclusion, Shoemaker's (1992) model is fundamentally flawed by two erroneous conclusions in regard to sheet flow versus channelized flow and by his misapplication of results on outburst-flood hydrographs. For this reason, I believe his results should be regarded with skepticism.

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