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Low-Froude-number stable flows past mountains(*)

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Summary. — A new approximate analysis is presented for stably stratified flows at low Froude number F past mountains of height H. In the "top" layer where the streamlines pass *above* the surface of the mountain, there is a perturbation flow. This approximately matches the lower flow in the "middle" 'horizontal' layer [M] in which the streamlines pass round the mountain in nearly horizontal planes, as in Drazin's (DRAZIN P. G., *On the steady flow of a fluid of variable density past an obstacle, Tellus*, **13** (1961) 239-251) model. The pressure associated with the diverging streamlines on the lee side of the summit layer flow drives the separated flow in the horizontal layer (which is not included in Drazin's model). This explains the vortical wake flow in experiments and in the "inviscid" computations of Smolarkiewicz and Rotunno (SMOLARKIEWICZ P. K. and ROTUNNO R., *Low Froude number flow past three-dimensional obstacles. Part I: Baroclinically generated lee vortices, J. Atmos. Sci.*, **46** (1989) 1154-1164). A method for estimating the height $H^{\mathrm{T}} \approx FH$ of the cut-off mountain is derived, as a function of upstream shear, mountain shape and other parameters. Recent laboratory experiments have confirmed how the *curvature* of the oncoming shear flow profile $(-\partial U^2/\partial z^2)$ can produce a significant reduction in the cut-off height H^{T} and in the distance downstream of the crest where the lee flow separates. This effect may reduce the wave drag of groups of mountains of similar height. The extension of the analysis to the movement of weak fronts past mountains is briefly described.

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1. – Introduction

The importance of mesoscale topographic effects has long been recognised by meteorologists (see, for example, the review by [1]). If we accept that the influence of a mountain or mountain range on the weather and general circulation is generally proportional to its size, then the importance of low Froude number flow effects, where Froude number is inversely proportional to mountain height (see below), can readily be appreciated. Under general atmospheric conditions, mountains over about 1 km in height are frequently associated with low Froude number flows, while the flow associated with mountains above about 2 km in height is in the low Froude number regime most of the time. But only recently have low Froude number effects begun to be represented in forecasting models (*e.g.* [2]), though there are some outstanding questions about the effects of finite Rossby numbers in these flows [3].

We consider steady, stably stratified flow (buoyancy frequency profile $N(z)$, with representative value N_0, mean flow profile $U(z)$ in the x-direction, with a representative value of velocity at the hill top of U_0) past an isolated three-dimensional mountain of height H and half-width L under the conditions that the Froude number is small ($F = U_0/N_0H \ll 1$) and that the Reynolds number is large ($R_e = U_0L/\nu \gg 1$, ν= kinematic viscosity). The basic structure of the flow, as confirmed by laboratory experiments and numerical simulations, is described in Sect. **2** below, together with some theoretical considerations which have been aired previously. In Sect. **3** a partial explanation of the shape of the wake as observed experimentally by Sysoeva and Chashechkin [4] is put forward. In Sect. **4** linear theory is used to gain a quantitative result connecting changes in the curvature of the mean flow profile with details of the structure of the flow. Section **5** summarises unpublished work of Greenslade, Hunt and Mobbs [5] in which the model is extended to allow for a horizontally inhomogeneous flow, as occurs in the atmosphere at a front. Our conclusions are presented in Sect. **6**.

2. – Structure of the flow

We examine the geophysically relevant parameter regime in which the *Froude number* is small

$$F \equiv \frac{U_0}{N_0H} \ll 1 \tag{1}$$

and the *Reynolds number* is large

$$R_e \equiv \frac{U_0L}{\nu} \gg 1. \tag{2}$$

A mathematically rigorous solution has not been achieved at the present time, nor is one likely to be. Furthermore, no approximate solution exists which is self-consistent and which explains all the observed features of the model (though the equations of motion have been integrated numerically). Our aim here is to review progress towards a suitable approximate solution.

The division of the flow into three main regions has been described by Greenslade [6, 7], see fig. 1. In the middle region [M], which extends over most of the depth of the mountain, air must flow *around* the mountain. Only in the top (or summit) region [T] (and above) does the flow rise over the mountain. A third region, the base region [B], lies beneath [M] so here again the air must flow around the mountain. (In exceptional model

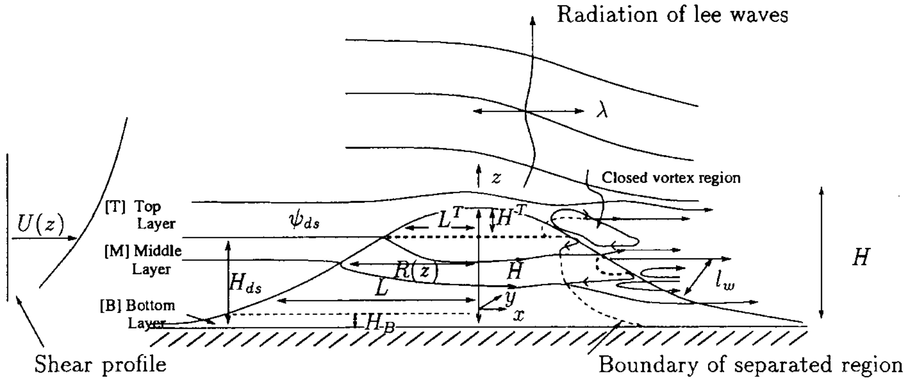

Fig. 1. – Schematic of upwind/side view of low Froude number flow over three-dimensional mountain.

cases such as a hemispherical mountain, [B] effectively does not exist.) The distinguishing property of both [T] and [B] is that the weak vertical motions are dynamically significant within these regions, whereas in [M] they have a second-order effect. This basic structure has been observed in laboratory experiments [4, 8–11], field experiments [12] and numerical simulations [13–16].

Note that it is possible to define a unique stream surface (isentropic surface) which separates [T] from [M]. The elevation of this surface far upstream from the obstacle (where it is horizontal) is known as the *dividing streamline height*, H_{ds}. Then, we can define $H^{\mathrm{T}} = H - H_{\mathrm{ds}}$ as a measure of the depth of [T].

As far as the flow field is concerned, Drazin's [17] theory gives an apparently self-consistent result for [M], comprising layerwise horizontal potential flow at leading order in F. This flow has only been realised in inviscid numerical simulations (for very small Froude numbers: $F < 0.1$ [14]). In the experiments and the simulations for $F \gtrsim 0.2$ there is a significant departure from the theory, because wakes form in the lee of the mountain, where the flow is dominated by a pair of vertically oriented eddies. Whether the eddies in the high Reynolds number laboratory experiments owe their existence to the same physical processes as the eddies in the inviscid numerical simulations, *i.e.* whether the limit $R_e \to \infty$ is singular, is an open question. Furthermore, it should be remembered that numerical simulations without explicit viscosity must in fact be dissipative to a degree determined by the accuracy of the numerical scheme.

For the flow in [T] Sheppard [18] used a physical argument based on a balance between kinetic and potential energy to indicate that, for an atmosphere of uniform properties (mean flow speed U_0 and buoyancy frequency N_0),

$$H^{\mathrm{T}} = U_0/N_0. \tag{3}$$

Reference [10] published the results of laboratory experiments confirming Sheppard's formula as a good approximation in practice for roughly axisymmetric mountains. Reference [10] also gave an extended version of Sheppard's formula for non-uniform atmospheres. However, it was pointed out by Smith [19] that Sheppard's argument is invalid because the kinetic energy is not zero at the hill top which is assumed by his argument. The theory of Greenslade [6] implies that the flow here is governed by nonlinear equations and confirms the order of magnitude estimate of Drazin [17] that

$$H^{\mathrm{T}} = O(U_0/N_0). \tag{4}$$

Similarly, the order of magnitude of the depth of [B] is

$$H^{\mathrm{B}} = O(U_0/N_0), \tag{5}$$

see ref. [6]. References [3] and [9] suggested that in the far field the influence of [T] is felt principally as upward and downward propagating internal gravity waves and there is some support for this idea from the results of previous laboratory experiments and numerical simulations. This paper proposes a different mechanism for the near field. Furthermore the experiments and simulations suggest that the flow in [T] is at least qualitatively similar to linear flow past an isolated mountain at a Froude number near unity [20]. In Sect. 4 we adopt this notion as a working hypothesis.

We note here some of the implications of the matching between [T] and [M]. First, quasi-linear flow in [T] has a pressure field with an asymmetric structure (positive anomaly upstream and negative anomaly downstream). This is inconsistent with Drazin's solution in [M] immediately below [T], for which the pressure field is fore-aft symmetric. However, the pressure field driven by a separated flow in [M] (*e.g.*, one with lee eddies) is approximately consistent with such a flow in [T] (Smith [20] also suggested this model for [T] but he did not discuss its matching with [M]). Second, the baroclinically generated (second order) vertical vorticity component associated with linear flow in [T] is consistent with the sense of circulation of the lee eddies in [M]; this idea has been pursued to its logical conclusion in [2] where the existence of the lee eddies is attributed to this baroclinically generated vorticity. Third, note the velocity shear at the interface between [T] and [M] when a recirculating wake is present. On the centreplane $y = 0$ in the lee of the mountain the flow is in the negative x-direction in the wake in [M], while in [T] the flow is in the positive x-direction. The associated shear $(\partial u/\partial z)$ dictates that there is strong vorticity in the positive y-direction at the interface between [T] and [M]. This is consistent with the presence of recirculating zones and cowhorn eddies which have been observed in laboratory flows at low Froude number (*e.g.*, [8]). See subsect. 4·2 and fig. 5.

Although as Greenslade [6] has shown, the flow in [B] is formally governed by the same nonlinear equations as those which govern the flow in [T], observations suggest that the dominant behaviour of the flow in [B] outside the wake region is similar (for a hill as in fig. 1) to that of linearised flow over a gently sloping hill. However, in the wake region the flow in [B] appears to be controlled by the wake flow in [M].

3. – Wake structure in [M]

Sysoeva and Chashechkin [4] have observed in laboratory experiments that the wake in the lee of a sphere at low Froude number has an approximately rectangular cross-section (sufficiently close to the obstacle). In fig. 2 we assume that the concept applies to a hill. This observation contradicts earlier intuitive concepts of the wake in region [M] based on assuming similarity with wakes in unstratified flows in which the wake cross-section closely conforms to the shape of the obstacle at every level (*i.e.* it is circular in the case of a spherical obstacle).

The underlying reason has not been established; nor have the limits of this approximation been explored. Some preliminary results obtained in a very recent separate experimental study [21] have shown that, as expected theoretically, vortices shed at an angle to the vertical tend to become vertical within a distance downstream of the body that is very small compared with the radius L of the body when $F \ll 1$. However, they have a finite vertical length that is comparable to the width of the wake. So if $H/L \gtrsim 1$ the vortices are decorrelated and the wake sides are not vertical. But if $(H/L) \lesssim F \ll 1$ the sides of the wake B_w are approximately vertical and the width $(2y_\mathrm{s})$ is comparable with the width $(2L^\mathrm{T})$ of the top layer. This is consistent with the detailed wake and vortex shedding experiments of Lin *et al.* [11] and of Professor Castro (unpublished).

We apply this assumption to an axisymmetric mountain with radius $R(z)$, for which the locations of separation are defined by $(x_\mathrm{s}, \pm y_\mathrm{s})$ at each height z. Let $x_\mathrm{s} = -R(z)\cos\theta_\mathrm{s}$, $y_\mathrm{s} = R(z)\sin\theta_\mathrm{s}$, where θ_s is the angle measured from the upstream stagnation point and our assumption implies that $\partial y_\mathrm{s}/\partial z \simeq 0$. We can now derive the value of y_s as F varies and how x_s varies with z, by considering separation at the dividing streamline height H_ds, where the local radius of the horizontal contour is $R_\mathrm{ds}(\simeq L^\mathrm{T})$. Consider the top of the middle layer at $z = H_\mathrm{ds}$, where

(6) $$y_\mathrm{s} \approx R_\mathrm{ds}\sin\theta_\mathrm{s}\,.$$

From the typical shape of a rounded mountain top, R_ds is related geometrically to H_ds by

(7) $$\left(\frac{R_\mathrm{ds}}{L}\right) \approx \left(\frac{H - H_\mathrm{ds}}{H}\right)^{1/P}, \quad \text{where } P = 2, 4 \ldots .$$

Equation (7) corresponds to the model shape of most idealised laboratory experiments. From (6) and (7), it follows that

(8) $$y_s \propto LF^{1/P}; \qquad x_s \propto \frac{1}{z^{1/P}} \qquad (\text{as } z \to 0),$$

see [9]. It follows from (8) that if $P = 2$, $y_\mathrm{s} \lesssim L/4$, for $F < 0.05$. Thus, the width of the separation is defined by the summit layer and is only really small (say $L/4$, and the Strouhal number of vortex shedding rises) for very small values of $F(\lesssim 0.05)$ and if the slope H/L is also small. But if $F = 0.2$, the wake width $y_\mathrm{s} \sim l_\mathrm{w}$ is of order L. The former case is consistent with the numerical results of [14, 15] and the latter with the experiments of [9].

4. – Approximate solution in [T]

4·1. *Matching solutions for* [T] *and* [M] *layers*. – Since from (4) the height H^T of the top layer [T] is much less than its radius L^T, linear theory can provide an approximate estimate (as proposed by [20, 22]). Assuming that the pressure is continuous at the dividing height H_ds, then the velocity fields in the layers [T] and [M] can be approximately modelled.

H^T can be estimated by matching the maximum perturbation pressure Δp_mx on the upwind face above and below this level. Below the level in [M]

(9) $$\Delta p_\mathrm{mx} = \Delta p_{\mathrm{stag[M]}} \simeq \frac{1}{2}\rho U^2,$$

(where $\triangle p = p - p_0$, and p_0 is the background hydrostatic pressure).

Since for rounded hills the length L^T is much greater than the height H^T, then from (4) the Froude number (based on the length scale L^T of the 'cut-off' mountain in [T]) is very small, *i.e.*

(10) $$U_0/(L^\mathrm{T}N) \ll 1.$$

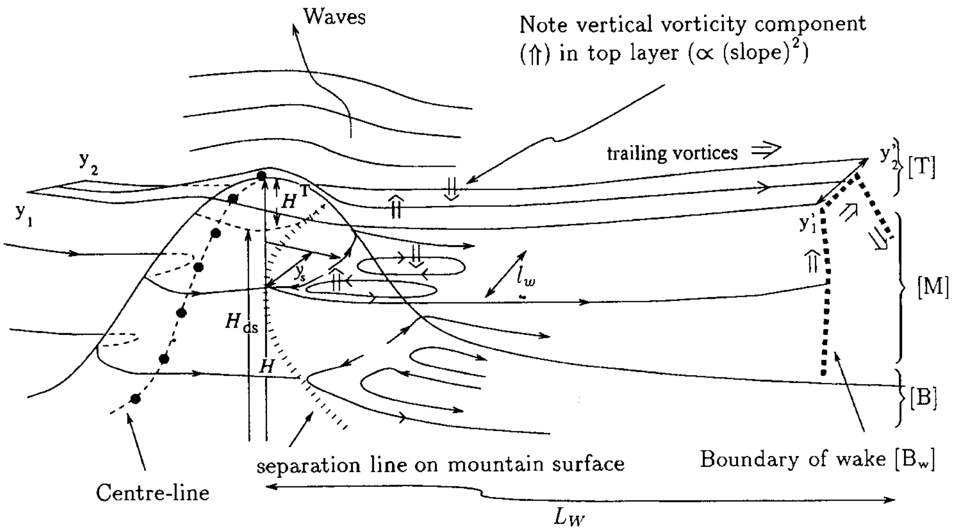

Fig. 2. – Schematic of side view/rear view.

Applying linear theory in [T], then along the typical streamline, lying above ψ_{ds} in fig. 1,

$$\Delta p \simeq \sigma_p \left(\frac{H^{\mathrm{T}} N}{U_0}\right) U_0^2, \tag{11}$$

where $H^{\mathrm{T}} \simeq H - H_{\mathrm{ds}}$ and σ_p is an $O(1)$ coefficient (or the normalised value of Δp). Thence, matching (10) and (11) at the upstream stagnation point where $\sigma_p = \sigma_{p_{\mathrm{mx}}}$ leads to

$$H^{\mathrm{T}} = \alpha_{\mathrm{T}} U/N, \tag{12}$$

where the coefficient

$$\alpha_{\mathrm{T}} \simeq \frac{1}{2\sigma_{p_{\mathrm{mx}}}}. \tag{13}$$

When (10) is satisfied, the coefficient $\sigma_{p_{\mathrm{mx}}}$ in (13) is approximately independent of $H^{\mathrm{T}}/L^{\mathrm{T}}$ for uniform flow over a mountain having a *given* shape and also varies little with shape. As H^{T} varies, the shape of the cut-off mountain varies, and therefore $\sigma_{p_{\mathrm{mx}}}$ is, in principle, a weak function of $H^{\mathrm{T}}/L^{\mathrm{T}}$. Linear theory shows that $\sigma_{p_{\mathrm{mx}}} \simeq 1/3$ for simple axisymmetric shapes, leading to the approximate theoretical value of

$$\alpha_{\mathrm{T}} \simeq 1.5\,. \tag{14}$$

Since for 2D *rounded* mountains, from equation (3.20) of [21]

$$\sigma_{p_{\mathrm{mx}}} = 0.7, \qquad \text{it follows that} \qquad \alpha_{\mathrm{T}} \simeq 0.7, \tag{15}$$

while for 2D triangular mountains (of any small angle, using the equation (13) in [22])

$$\sigma_{p_{\mathrm{mx}}} = \frac{2\ln 2}{\pi} = 0.44, \qquad \text{it follows that} \qquad \alpha_{\mathrm{T}} \simeq 1.1. \tag{16}$$

Experiments (*e.g.*, [10, 22]) confirm the general form of (12) and its insensitivity to slope. However, the experimental value of the coefficient is less than the theoretical prediction of (14), $\alpha_{\mathrm{T}} \simeq 1.0 \pm 0.3$, at least for mountains with moderate slopes ($H/L \lesssim 0.5$). Since $\sigma_{p_{\mathrm{mx}}}$ is found experimentally to increase both for rather large and for very low slopes, in these cases α_{T} may be even lower.

This approximate model enables α_{T} to be calculated approximately and thence the wave drag caused by the "cut-off" mountain (see [7]). The approximation involves incomplete matching, because the upper pressure, derived by assuming linear flow in [T] past a mountain over a flat surface ($z = H_{\mathrm{ds}}$), does not equate everywhere on the interface between [T] and [M] with the lower pressure in the horizontal free-streamline flow in [M]. Figure 3*a*) shows how the linear pressure decreases from the upwind to the downwind slopes as the flow accelerates, while fig. 3*b*) shows the classical form of the pressure distribution around a cylinder. Note that they are both asymmetric. For an exact match the dividing streamline (*e.g.*, ψ_{ds}) starting at height $z = H_{\mathrm{ds}}$, far upstream, should dip downwards (as fig. 1 shows, following [9]). A nonlinear Galerkin calculation might be constructed to obtain a numerical fully consistent nonlinear solution for the [T] and [M] layers.

Note that actual mountains have more complex shapes so that, when $F \ll 1$, there may be significant inertia-buoyancy layers (IB) at intermediate heights past the mountains, where $\mathrm{d}^2R/\mathrm{d}z^2 \gtrsim (U/N)$, *e.g.*, at both $H_{\mathrm{ds}}^{(1)}$ and $H_{\mathrm{ds}}^{(2)}$ in the mountain shown in fig. 4.

4·2. *Effect of curvature in the profile of the upstream velocity.* – The approximate matching of linear theory in [T] to horizontal motion in [M] enables us to calculate the effects of shear $U(z)$ in the incident profile. It is found that H_{ds} increases (and H^{T} decreases) by a small distance of order $(\mathrm{d}U/\mathrm{d}z)U/N^2$ (based on [8]) caused by the change in stagnation pressure in [M], and by a distance of order $(\mathrm{d}^2U/\mathrm{d}z^2)UH^{\mathrm{T}}/N^2$ caused by the effects of curvature of the profile on the flow in the top layer, following linear analysis [21].

The equation governing small perturbations to a unidirectional mean flow $U(z)$ with arbitrary shear and arbitrary buoyancy frequency profile $N(z)$ is, for the vertical velocity component,

$$\left\{U^2\frac{\partial^2}{\partial x^2}\nabla^2 + N^2\nabla_h^2 - UU''\frac{\partial^2}{\partial x^2}\right\} w = 0. \tag{17}$$

Fourier transforming from the physical (x, y)-plane to the (k, l)-plane gives

$$\left\{\frac{\partial^2}{\partial z^2} + \left(\frac{N^2}{U^2k^2} - 1\right)(k^2 + l^2) - \frac{U''}{U}\right\} \tilde{w} = 0. \tag{18}$$

From (18) the vertical scale L_z corresponding to a particular Fourier mode (k, l) is (of order)

$$L_z = \left\{\left(\frac{N^2}{U^2k^2} - 1\right)(k^2 + l^2) - \frac{U''}{U}\right\}^{-1/2}. \tag{19}$$

The exact spectrum of modes which are forced depends on the shape of the mountain. However, the dominant modes correspond to those (k, l)-values that are commensurate with the horizontal length scale of that part of the mountain in [T]. For a mountain with a rounded top this length scale is $F^{1/2}L$ [6], so the dominant modes have a wave number $\sqrt{k^2 + l^2} \sim 1/(F^{1/2}L)$ (smaller values of k and l are energetically important, but not

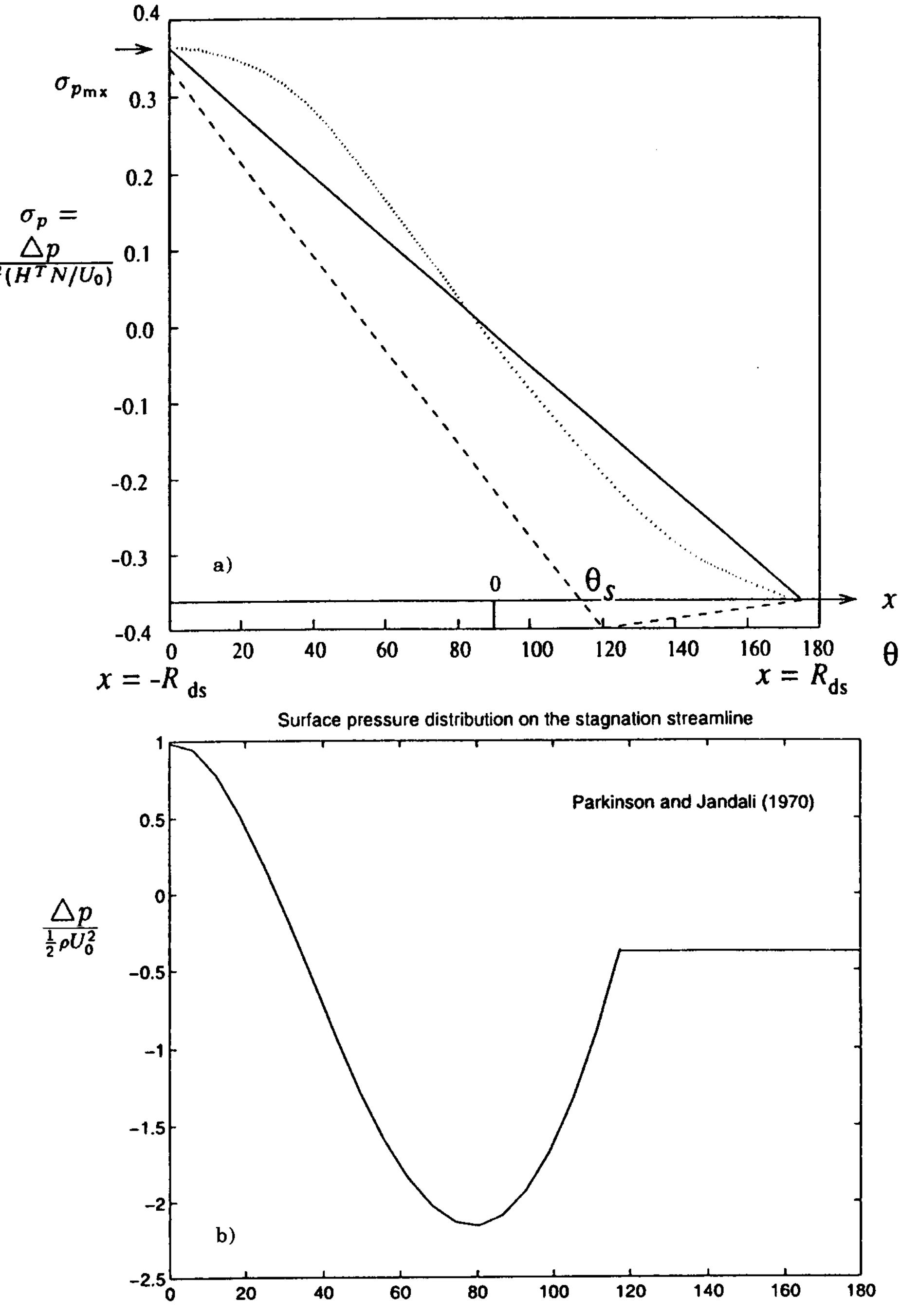

Fig. 3. – *a*) Perturbation pressure along $\psi_{\rm ds}$ as a function of the angle θ and streamwise distance x, where the dotted line is σ_p on $y = 0$; the solid line $\sigma_p(\theta)$ on $z = H_{\rm ds}$, based on Linear Theory [22] and a speculative curve (dashed line) $\sigma_p(\theta)$ if $z(\psi_{\rm ds})$ dips below $z = H_{\rm ds}$. $R_{\rm ds}$ is the radius of the hill at the height $z = H_{\rm ds}$ and $\theta_{\rm s}$ is the separation angle. *b*) Typical surface pressure in [M] at a fixed height for $\theta_{\rm s} = 120°$ for separated flow around a circuar cylinder.

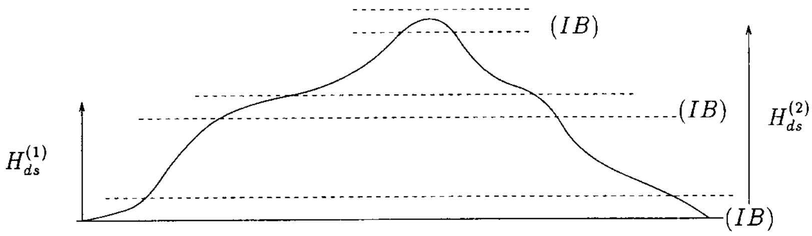

Fig. 4. – A typical mountain at low F showing the inertia-buoyancy layers each with thickness of order U/N.

larger values). Note then that the factor $(N^2/U^2k^2 - 1)$ approximates to N^2/U^2k^2 under the condition $F \ll 1$ (*i.e.* the dominant modes are hydrostatic). The predicted vertical scale then becomes

$$L_z \simeq \frac{U}{N}\left\{\frac{k^2+l^2}{k^2} - \frac{UU''}{N^2}\right\}^{-1/2}. \tag{20}$$

For a "typical" (k,l)-mode, $(k^2+l^2)/k^2$ is of order unity and we can absorb this term into α_{T}, resulting in the prediction

$$H^{\mathrm{T}} = \alpha_{\mathrm{T}} L_z, \tag{21}$$

where

$$L_z = \frac{U}{N}\left\{1 - \beta\frac{UU''}{N^2}\right\}^{-1/2}, \tag{22}$$

where $\beta = k^2/(k^2+l^2)$ is an order unity, positive constant. Thus:

$$U'' < 0 \Rightarrow L_z < \frac{U}{N}, \tag{23}$$

$$U'' > 0 \Rightarrow L_z > \frac{U}{N}, \tag{24}$$

i.e. H^{T} decreases with positive curvature of the mean flow ($U'' < 0$) and increases with negative curvature of the mean flow ($U'' > 0$).

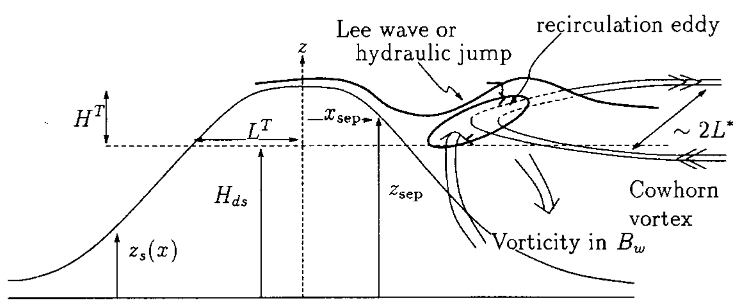

Fig. 5. – Lee side separation and wake structure in [T] above H_{ds} (mountain with moderate slope).

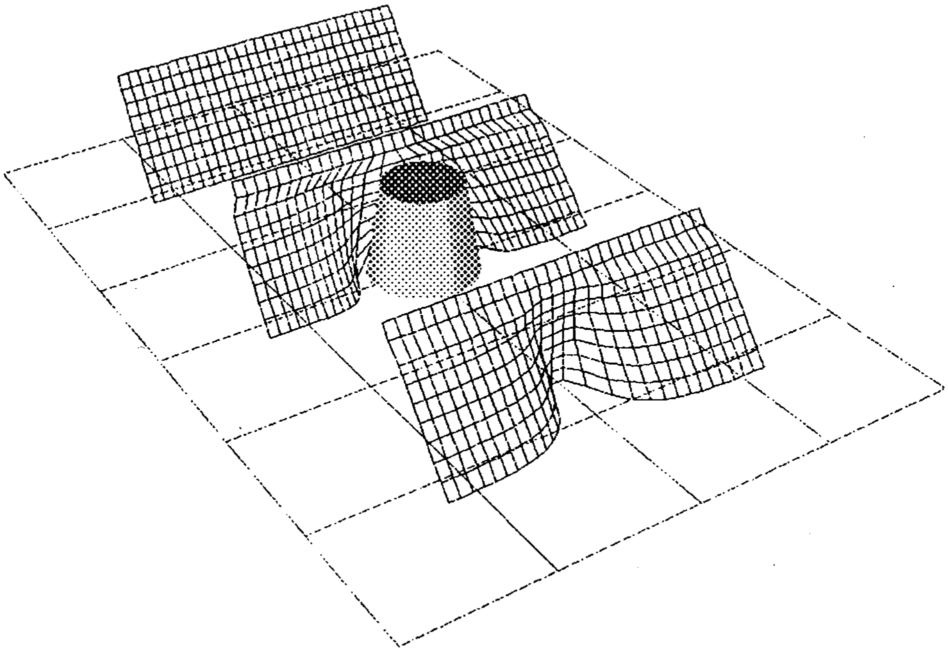

Fig. 6. – Transformation of a weak cold front induced by isolated orography at low Froude number.

For $F \gtrsim 0.1$, experiments show that the nonlinear effects in the boundary layer on the lee slope cause the flow to separate, to generate horizontal vorticity and cause lee waves to form at a position a distance $x_{\mathrm{sep}} \simeq \pi U/N$ from the top of the mountain, see fig. 5. Following the same argument as in sect. 4, it follows that the curvature of the incident profile effectively changes the value of N estimating x_{sep}. Reference [22] shows that x_{sep} is reduced by increasing the value of $(-\mathrm{d}^2U/\mathrm{d}z^2)$, according to

$$x_{\mathrm{sep}} \simeq \pi L_z \,. \tag{25}$$

The results (21) and (25) give some indication of the effects of *grouping of mountains*; they suggest that the wakes of upstream mountains of *about the same height* increase $|U''|$ in the approach flow of downstream mountains. This effect may therefore significantly reduce L_z and thence the wave drag of the top layer of these mountains as suggested by the experiment of [23].

5. – Fronts and orography

Greenslade, Hunt and Mobbs [5] have proposed a model for the influence of a mountain range on a travelling weak front based on the low Froude number flow theories described above. Their work which complements previous theoretical studies of front-orography interaction for higher Froude number (see [24] and references therein) is consistent with observations that weak fronts are strongly deformed under low Froude number (blocked) flow conditions, even far downwind of the orography. Furthermore, cold fronts are caused to overturn because the orographic blocking occurs further upwind at lower levels (where the mountain is wider) and nearer the mountain at higher levels. Overturning initiates a sequence of local convective instability, mixing and frontolysis, the net result of which is the formation of a transformed front downwind of the mountain. The location and timescale of overturning are identified for selected model orographies. A number of possible causes for the observed asymmetry of the deformation have been identified (*e.g.*, circulation caused by asymmetry of the mountain shape and the barrier jet caused by Coriolis effects).

Figure 6 (after [5]) shows a simulation of the transformation of a cold front induced by isolated orography in the form of a truncated cone. (For simplicity $F = 0$ and wake effects are neglected.) The figure shows a perspective view of the orography with three successive positions of a cold front during the transformation process. Note that the vertical scale in the figure is exaggerated one hundredfold (the typical slope of a cold front is 0.01). The initial position of the front, at left, is simply a planar, sloping surface. As the front moves towards the orography (from left to right), it begins to overturn close to the upwind slope of the orography (middle position). With convective instability and mixing processes taken into account, the overturning portion of the front transforms to a new position downwind of the orography (right). The net result of the transformation process is a "dent" in the front corresponding to the position where the orography passed through it. The size and shape of the dent correlate closely with the size and shape of the orography.

6. – Conclusions

In this paper we have described some recent concepts for studying the airflow past an isolated mountain at low Froude number. Although there is as yet no complete mathematical solution, there is some progress in fitting together a consistent description of the basic elements of the flow. Simple theories lead to useful estimates of a number of the observed features of the flow, including the effect of upward effects such as curvature in the shear flow and horizontal potential temperature (density) variations such as those associated with fronts. In the latter case it is shown that a weak cold front approaching a mountain always overturns, thus explaining the well-established observations of mixing and precipitation around the mountain and a transformed frontal surface downwind of the mountain, with a dent in the front marking the influence of the orography for a long distance downward. The asymmetric aspects that are observed in practice are the subject of further study.

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REFERENCES

[1] Smith R. B., *The influence of mountains on the atmosphere, Adv. Geophys.*, **21** (1979) 87-230.

[2] Lott F. and Miller M., *A new sub-grid scale orographic drag parametrization: Its formulation and testing, Q. J. R. Meteorol. Soc.*, **123** (1997) 101-127.

[3] Newley T. M. J., Pearson H. J. and Hunt J. C. R., *Stably stratified rotating flow through a group of obstacles, Geophys. Astrophys. Fluid Dyn.*, **58** (1991) 147-171.

[4] Sysoeva E. Y. and Chashechkin Y. D., *Spatial structure of a wake behind a sphere in a stratified liquid, J. Appl. Mech. Tech. Phys.*, **5** (1988) 655-660.

[5] GREENSLADE M. D., HUNT J. C. R. and MOBBS S. D., *An idealised model for the strong interaction between a weak front and mesoscale orography*, unpublished note; preprint available on request.
[6] GREENSLADE M. D., *Strongly stratified airflow over and around mountains*, PhD thesis, University of Leeds, 1992.
[7] GREENSLADE M. D., *Strongly stratified airflow over and around mountains* in *Stably Stratified Flows: Flow and Dispersion over Topography, Proceedings of the IV Conference on Stably Stratified Flows, September, 1992*, edited by I. P. CASTRO and N. J. ROCKLIFF (Clarendon Press, Oxford) 1995.
[8] BRIGHTON P. W. M., *Strongly stratified flow past three-dimensional obstacles, Q. J. R. Meteorol. Soc.*, **104** (1978) 289-307.
[9] HUNT J. C. R. and SNYDER W. H., *Experiments on stably and neutrally stratified flow over a model three-dimensional hill, J. Fuid Mech.*, **96** (1980) 671-704.
[10] SNYDER W. H., THOMPSON R. S., ESKRIDGE R. E., LAWSON R. E., CASTRO I. P., LEE J. T., HUNT J. C. R. and OGAWA Y., *The structure of strongly stratified flow over hills: Dividing-streamline concept, J. Fluid Mech.*, **152** (1985) 249-288.
[11] LIN Q., LINDBERG W. R., BOYER D. L. and FERNANDO H. J. S., *Stratified flow past a sphere, J. Fluid Mech.*, **240** (1992) 315-354.
[12] TAMPIERI F. and HUNT J. C. R., *Stratified flow and wave motion near mountain tops*, in *Proceedings of the Conference on the Scientific Results of the Alpine Experiment* (*ALPEX*). GARP publications series No. 27, pp. 359-369 (1986).
[13] HANAZAKI H., *A numerical study of three-dimensional stratified flow past a sphere, J. Fluid Mech.*, **192** (1988) 393-419.
[14] SMOLARKIEWICZ P. K. and ROTUNNO R., *Low Froude number flow past three-dimensional obstacles. Part I: Baroclinically generated lee vortices, J. Atmos. Sci.*, **46** (1989) 1154-1164.
[15] SMOLARKIEWICZ P. K. and ROTUNNO R., *Low Froude number flow past three-dimensional obstacles. Part II: Upwind flow reversal zone, J. Atmos. Sci.*, **47** (1990) 1498-1511.
[16] SUZUKI M. and KUWAHARA K., *Stratified flow past a bell-shaped hill, Fluid Dyn. Res.*, **9** (1992) 1-18.
[17] DRAZIN P. G., *On the steady flow of a fluid of variable density past an obstacle, Tellus*, **13** (1961) 239-251.
[18] SHEPPARD P. A., *Airflow over mountains, Q. J. R. Meteorol. Soc.*, **82** (1956) 528-529.
[19] SMITH R. B., *Mountain-induced stagnation points in hydrostatic flow, Tellus A*, **41** (1989) 270-274.
[20] SMITH R. B., *Linear theory of stratified hydrostatic flow past an isolated mountain, Tellus*, **32** (1980) 348-364.
[21] HUNT J. C. R. and FERNANDO H. J. S., *Separated flow around bluff bodies at low Froude number: votex shedding and estimates of drag, Proc. V IMA Conference on Stably Stratified Flows, Dundee, Sept. 1996*, edited by P. A. DAVIES (Clarendon Press, Oxford) 1988.
[22] HUNT J. C. R., RICHARDS K. J. and BRIGHTON P. W. M., *Stably stratified shear flow over low hills, Q. J. R. Meteorol. Soc.*, **114** (1988) 859-886.
[23] FENG Y., *Stably stratified shear flow over complex terrain*, PhD dissertation, University of Cambridge, 1995.
[24] EGGER J. and HOINKA K. P., *Fronts and orography, Meteorol. Atmos. Phys.*, **48** (1992) 3-36.