On the Problem of Violent Valley Winds

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ABSTRACT

Observational results of a one-month mesoscale experiment in a valley are used to emphasize the prominent part played by an inversion layer in air flow dynamics. A model based on the analogy between shallow water flow and air flow beneath an inversion is applied. Using the actual topography of the valley, theoretical arguments for the appearance of violent winds are presented and the model results are compared with the in-situ measurements. The qualitatively good agreement for the wind speed and the height of the inversion layer allows for further confidence in the model predictions. For example, the existence of a sharp transition zone, or jump, with a strong horizontal gradient of wind speed and temperature present in the model is consistent with aircraft measurements.

1. Introduction

An integrated system of meteorological measurements using simultaneously radio soundings, radar winds, pilot balloons and aircraft had been set up in a large valley (the Rhône Valley in France) for a one-month mesoscale experiment. The analysis of the measurements showed striking consistency in the data, in particular for the case of violent wind conditions in the southern part of the valley.

This violent wind, called the ‘Mistral’, is well known in France (see, e.g., Queneu, 1974) but has been inadequately explained up to now, partly because the corresponding studies (e.g., Galizi, 1952) were hampered by lack of information. The experimental program, called “Campagne Vallée du Rhône 77”, brings out the following new elements for the study of the violent valley wind:

1) A high spatial resolution from the synoptic scale, provided by the French meteorological network, down to the mesoscale, provided by the high number of soundings throughout the valley, and even to the microscale, provided by the aircraft measurements.

2) A fine temporal resolution achieved through the high frequency of daily soundings, allowing for a description of the diurnal cycle.

It became apparent during the experiment that strong northerly winds in the southern part of the valley were related to the existence of a temperature inversion layer. In order to depict the observed flow we consequently adopt a two-layer model with an underlying mean flow capped by an inversion layer. Taking into account the actual topography of the valley, this model leads to a bulk structure of the observed violent winds which compares favorably with the in-situ measurements, with respect to the wind speed as well as the height of the inversion layer. The most interesting result of this comparison is that it confirms the existence of a transition zone characterized by a strong horizontal gradient of wind speed and temperature. This feature was in fact present in the aircraft measurements but was at a scale too small to be detected by the other measurement systems.

2. The experimental program “Campagne Vallée du Rhône 77”

Fig. 1 shows a map of the Rhône Valley established from the relatively smooth topography used for numerical modeling. This valley is located in southeastern France, between two large mountain ranges, the Massif Central (highest elevation, 1885 m) and the Alps (highest elevation, Mont Blanc, 4807 m). The experimental program was devoted to the study of a large area of ~600 km from north to south and 400 km from east to west. Seven radiosounding and radar-wind stations were taking measurement of pressure, temperature, moisture and wind speed and five pilot-balcony stations were measuring only wind speed (see Fig. 1).

In addition, two meteorological aircraft were operated (Cessna and Aerocommander). The flight plans were as follows:

1) Northern flight plan: four flight levels (500, 1000, 1500 and 2000 m) giving a vertical cross section of ~40 km along the axis of the valley and centered on Valence; five flight levels (3000, 2000, 1500, 1000 and 600 m) giving a vertical cross section...
of \( \sim 40 \) km at right angles to the valley axis and located \( \sim 25 \) km north of Valence.

2) Southern flight plan: three flight levels (1000, 1500 and 2000 m) giving a vertical cross section of \( \sim 50 \) km along the valley axis and centered on Montelimar; five flight levels (3000, 2000, 1500, 1000 and 500 m) giving a vertical cross section of \( \sim 40 \) km at right angles to the valley axis and located \( \sim 20 \) km north of Orange.

The program took place between 17 November and 7 December 1977. The resulting data set consists of observations from 437 radio soundings, 437 radar winds and 242 pilot balloons, as well as from 50 h of flight for the Aerocommander and 60 h for the Cessna.

3. The north wind situations in the Rhône Valley

Six north wind situations were documented during the experimental program: 18, 19, 23 and 27 November and 2 and 3 December. The case of 3 December is somewhat unusual in the sense that fewer soundings were taken.

For all of these six cases both the wind speed and temperature observations were quite coherent throughout the day, even if the aircraft soundings sometimes show significant variations of an undulatory character on a smaller scale. Thus, to simplify the interpretation and modeling of the results, we have taken daily averages of the soundings. These averages are representative of the airflow properties in the Rhône Valley for these days.

The mean wind speed and potential temperature profiles are shown in Figs. 2a and 2b. It can be seen that there is a strong difference between the moderate winds in the north (Cuisery, Feurs, Satolas) and the violent winds in the south (Valence, Orange, Nîmes). One also observes a warming of the air from north to south with a lowering of the inversion layer.
The main features common to the north wind situations are that 1) the wind speed increases from Valence downstream; 2) when a violent wind blows in the southern part of the valley, its maximum speed is located at Orange; 3) there is a warming of the air from north to south; 4) the violent winds in the south of the valley occur mostly in the presence of an inversion layer; and 5) when an inversion layer is present, its base lowers by several hundred meters from north to south, causing a sharp decrease of the sublayer depth.

Fig. 3 shows the north–south potential temperature cross sections on 19 November from 0900 to 1000 GMT and on 23 November from 1230 to 1350 GMT, measured from the two aircraft. Also shown is the sign of the vertical component of velocity, but only one of the two aircraft could measure it. Both cross sections show strong contrast between the
northern and southern temperature field patterns. The isotherms upwind of Valence are quasi-horizontal. Downstream of Valence the isotherms lower by 200 or 300 m. Further downstream, south of Montelimar, the isotherms clearly show a wavy pattern.

From these experimental results one can distinguish between two kinds of flow in the valley: a tranquil streaming in the northern upstream section and a flow dominated by wave motions in the southern downstream section. The transition from one flow to the other is quite sharp and occurs near Montelimar. The wave which can be seen in the south is thought to be a stationary gravity wave, since it requires more than one h for the aircraft to traverse it. The depth h of the stable layer remained at less than 300 m on 19 and 23 November. The possibility of trapped waves can thus be eliminated since the corresponding wavelength $\lambda = 7.5e$ (Turner, 1973), e being the depth of the inversion layer, is much smaller than the observed wavelengths (see Table 1).

Fig. 2b. As in Fig. 2a for the north wind situations of 27 November and 2 and 3 December 1977.
Analysis of the measurements seems to prove that violent winds in the southern Rhône Valley are induced by wave forcing. In another way, the results show that the temperature inversion plays a prominent part in the appearance of this phenomenon. The only day (18 November) on which no temperature inversion layer was present is also the day when the wind acceleration was the weakest. Between Satolas and Orange the wind speed then only doubled, whereas it can quadruple when there is an inversion layer (e.g., on 2 December).

4. A model of air flow beneath an inversion layer

The theory of strong katabatic winds (Ball, 1956) seems to be the most convenient framework in which to interpret the violent winds occurring in the southern part of the Rhône Valley. The aim of this section is to recall the very specific aspects of a simplified theory by Arakawa (1968) for katabatic valley winds. This theory will be used as a theoretical framework for the interpretation of experimental observations.

a. Quasi two-dimensional stationary flow

We now consider a channel flow in which two incompressible fluids of different densities are separated by an interface, as shown in Fig. 4. The upper fluid is at rest while the depth h of the moving lower fluid is relatively shallow. We further assume that the effects of the earth’s rotation can be neglected and that the hydrostatic approximation is valid.

For greater simplicity, the flow is supposed to be
stationary and homogeneous in the vertical and transverse directions \( z \) and \( y \), with no superimposed pressure gradient. The channel section is taken as being rectangular, with only gradual longitudinal variation of its width \( b \) and depth \( (H' - m) \) (Fig. 4).

Under these hypotheses and neglecting frictional forces, the continuity and momentum equations can be integrated along the transverse direction by assuming that the channel walls are material surfaces. One then has

\[
\frac{d}{dx}(abh) = 0, \tag{1}
\]

\[
u \frac{du}{dx} + g' \frac{dh}{dx} + g \frac{dm}{dx} = 0, \tag{2}
\]

where \( g' = g(\rho - \rho')/\rho \) is the reduced gravity, and \( \rho \) the density.

Eliminating \( dh/dx \) between Eqs. (1) and (2) yields

\[
(F^2 - 1) \frac{1}{u} \frac{du}{dx} = \frac{1}{b} \frac{db}{dx} - \frac{1}{h} \frac{dm}{dx}, \tag{3}
\]

while eliminating \( du/dx \) yields

\[
(F^2 - 1) \frac{1}{h} \frac{dh}{dx} = \frac{1}{b} \frac{db}{dx} - \frac{1}{h} \frac{F^2 db}{dx}. \tag{4}
\]

In Eqs. (3) and (4) the Froude number \( F = \frac{u^2}{g'h} \) is the ratio between the velocity of the fluid and the velocity of the gravity wave at the base of the layer.

In the following paragraphs we shall give several definitions concerning the local properties of the flow used. A flow region is called sub- (super-) critical when \( F \) is lower (greater) than 1. If \( F \) reaches locally the value 1, this flow region is called critical. The specific problem we want to deal with concerns a flow in which \( F \) can cross the critical value 1 at some particular point. It can be seen that the right-hand side of (3) and (4) are very unlikely to vanish in the general case. Thus, if \( F \) takes the value 1, this implies that \( |du/dx| \) and \( |dh/dx| \) become infinite. At such a particular point, the flow presents a sharp variation (jump) of both wind speed and layer depth.

It may nevertheless happen that

\[
\frac{1}{b} \frac{db}{dx} - \frac{1}{h} \frac{dm}{dx} = 0, \tag{5}
\]

at a point which will be called a "specific point". Then, at a specific point, taking into account the two local conditions \( F = 1 \) and Eq. (5) and by using Eq. (1), Eq. (3) can be differentiated with respect to \( x \), leading to

\[
\frac{3 du}{u dx} + \frac{2 db}{b dx} = \frac{1}{u} \frac{du}{dx},
\]

\[
= \frac{1}{b} \frac{d^2 b}{dx^2} - \frac{g'}{u^2} \frac{d^2 m}{dx^2} - \frac{2}{b^2} \left( \frac{db}{dx} \right)^2. \tag{6}
\]

Generally, the two sides of (6) are not equal to zero and \( du/dx \), as well as \( dF/dx \), are neither zero nor infinite. This implies that if the critical value \( F = 1 \) is reached at a specific point, the critical nature (sub or super) is different for the upstream and downstream regions of the flow on both sides of the specific point.

If (5) is fulfilled when \( F \neq 1 \), Eq. (3) implies that \( du/dx = 0 \). The wind speed is then an extremum at the specific point. In this case, if one further assumes that the channel walls are parallel \( (b = \text{constant}) \), the extrema for both wind speed and depth \( h \) occur at the specific point where \( dm/dx = 0 \), i.e., at a crest of the bottom of the valley. In the alternative case of a flat-bottomed valley \( (m = 0 \text{ everywhere}) \), the extremum for both wind speed and depth \( h \) occur at the specific point where \( db/dx = 0 \), i.e., at a constriction of the valley.

b. Conditions for continuous flow

We are now going to look for the conditions in which a stationary flow can be the solution of the above equations. We shall restrict our discussion to the special case \( m = 0 \). To do this, the relationship between the width of the channel and the wind speed when the critical flow occurs at the specific point must first be derived.

The suffix \( i \) will refer to the upstream flow, the suffix \( t \) to any point downstream. Integrate (1) and (2) from upwind to any location gives

\[
\begin{align*}
&u_i h_i b_i = u_h b_i = Q, \tag{7} \\
&u_i^2/2g' + h_i = u_t^2/2g' + h_t = E. \tag{8}
\end{align*}
\]

At a specific point \( c \) where it is further assumed that the critical value \( F_c = u_i^2/(g'h_i)^{1/2} = 1 \) is reached, one gets from Eqs. (7) and (8)

\[
F_i^2 - 3(F_i/B_c)^{2/3} + 2 = 0, \tag{9}
\]

in which, as previously, \( F_i \) represents the Froude number at the entry of the valley and \( B_c = b_c/b_i \) the ratio between the valley width \( b_c \) at the point where the critical flow occurs, and \( b_i \) at the entry. Relation (9) between \( B_c \) and \( F_i \) is shown in Fig. 5. At any point \( t \) in the valley, we can also write from Eqs. (7) and (8)

\[
\frac{1}{2} F_i^2 + 1 = \frac{1}{2} F_t^2 (A_t^2 B_t^2)^{-1} + A_t, \tag{10}
\]

in which \( A_t = h_t/h_i \) and \( B_t = b_t/b_c \). By subtracting Eq. (10) written at point \( t \) from Eq. (10) written at point \( c \) and by using Eq. (7) written at point \( c \), we have

\[
(1 - B^2) A^3 = A_i^3 - 3A_i + 2, \tag{11}
\]

where \( A_c/A_t = A \) and \( B_c/B_t = B \). Since \( A \) cannot be negative, Eq. (11) implies that

\[
1 - B^2 > 0 \quad \text{or} \quad B_i > B_c. \tag{12}
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<td>0.75</td>
<td>1.19</td>
<td>122</td>
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</table>
Table 1. (Continued)

<table>
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<tr>
<th>Stations</th>
<th>Hour (GMT)</th>
<th>$\theta'$ (K)</th>
<th>$\theta$ (K)</th>
<th>h (m)</th>
<th>$\delta$ (m s$^{-1}$)</th>
<th>$u_{cm}$ (m s$^{-1}$)</th>
<th>c</th>
<th>F</th>
<th>$h_1/h_i$</th>
<th>e (m)</th>
</tr>
</thead>
</table>

23 November 1977

Satolas 657 284.5 281.0 2095 6.82 9.5 15.9 0.43 1 152
Valence 654 289.0 284.0 2467 14.01 16.6 20.5 0.68 0.85 147
Orange 650 288.9 282.2 2050 13.19 20.5 21.9 0.6 1.02 602
Nimes 858 287.2 284.5 2243 7.32 14.0 14.38 0.51 1 259
Valence 858 287.4 280.5 1871 13.09 15.0 20.99 0.62 0.83 100
Orange 908 289.0 280.9 1354 20.00 25.5 19.29 1.04 0.6 801
Nimes 850 291.2 281.2 1808 13.23 17.5 25.10 0.53 0.81 494
Satolas 1140 287.5 285.8 2424 5.9 9.5 11.8 0.50 1 63
Valence 1158 291.5 281.4 1690 11.4 12.7 23.9 0.48 1.23 158
Orange Nimes 1125 287.9 282.0 1710 9.75 12.5 18.73 0.52 1.25 290
Satolas 1455 288.8 285.6 2295 4.65 7.5 15.79 0.29 1 294
Valence 1500 290.1 283.5 2007 12.60 15.7 21.16 0.60 0.87 152
Orange 1523 289.1 283.2 1844 18.94 25.5 15.62 1.21 0.80 160
Nimes 1450 290.1 283.9 1690 13.85 22.0 19.03 0.73 0.74 278

Eq. (12) states that if the flow becomes critical this can only occur at the specific point where the width of the channel is an absolute minimum. This happens only in the case where the Froude number at the entry of the valley is equal to the particular value determined from the geometry of the valley [see Eq. (9) and Fig. 5]. If the Froude number at the entry is not equal to this particular value, two cases must then be considered. First, if Eq. (10) has a positive real root, the flow properties become an extremum at the specific point [see Eqs. (3) and (4)]. On the other hand it may happen that Eq. (10) has only non-physical roots. This latter case will be discussed with more details in the next subsection.

c. Extension to super-critical flow and jump

What happens when the flow at the entry of the valley does not obey Eq. (9)? Such a case can occur, for example, when it is too speedy to become critical at the specific point. In this case it can be assumed (and this is confirmed by observation) that, instead of a continuous flow in the valley, the air is “blocked” by the narrow section and a sharp swelling zone occurs. The air is then slowed down and the height of the inversion layer increases. The Froude number consequently decreases, eventually reaching a value

FIG. 4. Air flow beneath an inversion layer in a rectangular channel. The base of the inversion is idealized as an interface between two fluids of different densities $\rho$ and $\rho'$.  

FIG. 5. Froude number $F_1$, at the entry of the valley as a function of the ratio $B_1$ of the width at the constriction to the width at the entry in the case of a flat-bottomed valley.
Fig. 6. Schematic side view of the air flow beneath an inversion layer in a flat-bottomed valley when the Froude number of the upstream flow is greater than the particular value leading to critical ($F = 1$) flow in the narrowest section. Dotted lines show the top view width of the valley.

corresponding to the conditions for which the above analysis would apply, i.e., for the flow to become critical at the specific point. Fig. 6 shows a schematic representation of such a phenomenon.

From now on we shall assume that, eventually due to a possible swelling zone, the flow at the entry obeys Eq. (9) and is subcritical ($F < 1$). The ratio $H = h/h_i$ between the current height of the inversion layer and the height of the inversion of the upstream flow, is related to the upstream Froude number $F_i$ by Eq. (10) which can be rewritten as

$$2H^3 - (F_i^2 + 2)H^2 + F_i^2/B_i^2 = 0. \quad (13)$$

This cubic equation has three real roots if $F_i$ remains less than or equal to the particular value for which $F = 1$ at the specific point (point c). Two roots are positive, the third is negative and non-physical. The current root is the one which can be deduced by continuity from the solution of (13) at $x = x_i$, which obviously is $H = 1 (h = h_i)$. If $F$ does not take the value 1, the current solution is symmetric on both sides of the specific point. If, at the specific point, $F$ takes the value 1, we have shown that the flow changes there from subcritical to supercritical. At the specific point, the two positive roots are equal and downstream of the specific point, the current root is the one which does not lead to a symmetric behavior. In this case, the height of the inversion layer decreases and the wind speed increases in the rest of the valley, downstream of the specific point. As the Froude number cannot reach infinity, the flow must again become sub-critical, even if there is no specific point downstream. One must then retain the assumption that the flow somewhere undergoes a sharp transition or a jump where $|du/dx| \to \infty$.

According to Lamb (1930), Ball (1956) or Stoker (1957), the height conditions on both sides of a stationary jump are given by

$$h_2 = \frac{1}{2}h_i[(1 + 8F_i^2)^{1/2} - 1], \quad (14)$$

where the indices 1 and 2 refer respectively to the upstream and downstream positions. Eq. (14) indicates that the magnitude of the jump is determined only by the upstream conditions. On the other hand, we shall show in Section 5b that the location of the jump is determined by the height difference between the inversion layer at the entry and at the exit of the valley. The greater this difference, the further from the specific point of the valley the location of maximum wind speed is.

Fig. 6 summarizes the above flow characteristics. In the upstream part of the valley, the Froude number is low (sub-critical) and the height of the inversion layer varies slowly. The flow then becomes critical in the narrowest section of the valley, downstream of which the height of the inversion layer decreases rapidly as the velocity increases. Downstream of that specific point, the flow is then super-critical. When the jump takes place, the height of the inversion layer suddenly increases and the wind speed drops sharply. Downstream of the jump, the flow becomes sub-critical again, up to the exit of the valley. It could then be shown from Eqs. (13) and (14) that the height of the inversion layer at the exit is less than that at the entry.

5. Observed airflow beneath an inversion layer in the Rhône Valley

We apply here the above theoretical model to the actual situation in the Rhône Valley, using the real topography, and compare its predictions with in-situ measurements.

a. Orography of the Rhône Valley and determination of the specific point

We used the topography of the southeastern part of France as provided by the Institut Géographique National. The topography of the area located approximately between 2 and 7.5°E and 42 and 46.5°N is given on a grid of 1 km mesh with 455 points in the east–west direction and 478 points in the north–south direction.

A series of horizontal cross sections with isopleths separated by 100 m from 500 up to 1500 m has been obtained and is shown in Fig. 7. It can first be noticed (see Fig. 8) that the ratio $b/b_i$ between the width at a point of the valley and the width at the entry, near Lyon, is more or less constant with height. Fig. 8 also shows that, whatever the altitude one is considering, the narrowest point of the valley is always located at the same place. From Figs. 7 and 8 the ratio $b/b_i$ can consequently be considered as being roughly constant with height.

The altitude of the valley floor has been determined every kilometer along the valley and small-scale variations have been filtered out. The maximum altitude, which is $\sim 530$ m, is observed half way between Lyon and Valence, 45 km north of the narrowest part of the valley, located near Valence. Up-
stream, at Lyon, the height of the valley floor is \(\sim 250\) m. The slope of the floor is weak (2.5\%) while the narrowing of the valley is quite rapid; \(b/b_0\) varies from 1 to 0.5 over 100 km.

We also need to determine the location of the specific point, i.e., where condition (5) is fulfilled. In fact, depending on the height of the inversion and on the choice of the altitude at which the valley width is determined, there can be several places where (5) is verified. Nevertheless, relation (5) is almost always satisfied between Valence and Montelimar.

The above features led us to consider the interpretation of experimental data within the framework of a flat-bottomed Rhône Valley, where the specific point is taken at the narrowest point \((b/b_0 = 0.49)\) 100 km south of Lyon, near Valence.

b. Phenomenological description of certain violent valley winds

Fig. 9 gives a complete phenomenological description of certain violent winds. The width of the valley \((b/b_0 = 0.49)\) at the specific point requires a Froude number at the entry (near Lyon) smaller than or equal to 0.28, for the analysis of Section 4 to be applied. Eq. (13) gives then the depth of the lower layer up to the specific point as shown by curve A in Fig. 9. The height of the inversion layer first decreases slowly for the first 70 km, then more rapidly when approaching the narrow section. The flow is critical at the specific point near Valence and becomes super-critical downstream (curve C), in which case the height of the inversion layer would decrease rapidly after the narrow section and then more slowly. At the same time, the Froude number increases rapidly from Valence to Nimes and the wind speed at Orange reaches five times its value at Lyon. A jump must then occur somewhere to reduce the Froude number to a value < 1. Relation (14) allows the drawing of curve D which gives the height of the inversion layer after such a jump. Taking these as a new initial condition it is possible to reuse Eq. (13) for the flow after the jump. Depending on the location where this jump takes place, the height \(h_D\) of the inversion layer at the exit of the valley will be different; thus, \(h_D/h_0 = 0.9\) is obtained if the jump
occurs at $J_1$ north of Orange (curve E), while $h_D/h_i = 0.7$ corresponds to a jump located at $J_2$ south of Orange (curve F). In fact, the height of the inversion layer at the exit of the valley is determined by synoptic-scale conditions, so that the location of the jump can be inferred from these conditions. The symmetrical solution of Eq. (13) is also shown in Fig. 9 (curve B), in which case the flow remains subcritical downstream of the specific point.

Several general statements may be deduced from the above arguments.

- The height of the inversion layer varies rapidly in the narrow section of the valley, close to the specific point.
- The Froude number increases rapidly downstream of the narrow section, which explains the appearance of violent winds.
- A height of the inversion layer much lower at the exit than at the entry of the valley corresponds to a jump occurring far downwind, eventually south of Nimes. We can thus expect very violent winds extending up to the coast. On the other hand, a small height difference between the exit and the entry corresponds to a jump occurring immediately after the narrow section, in which case no violent winds are
to be expected in the valley. Despite these markedly different behaviors, there is no basic dynamical difference between these two flows.

c. Comparison with experimental data

It still remains to check these results against the measurements taken during the Campagne Vallée du Rhône 77.

As stated in Section 3, six north-wind situations have been documented. The case of 18 November is excluded as there was no marked temperature inversion (see Fig. 2a), while the case of 3 December must be eliminated because of further complications due to superimposed fluxes. Table I gives the results for the four remaining days at each station. Some of these results are also shown in Fig. 9. For the sake of simplicity only half of these data have been retained in the present analysis, namely those taken at about 0700 LT (1200 GMT). These two periods also correspond to the aircraft flights. The choice for 23 November is different because several soundings at Orange are missing.

The ratio \( h/h_i \) and the Froude Number at each station are plotted in Fig. 9. The ratio \( h/h_i \) between the height of the inversion layer at a specific station and the one at Satolas takes values in rather good agreement with those given by the theory. On the other hand, the Froude number is almost always greater than the one predicted by the theory, although remaining generally within an acceptable margin.

Two series of measurements are particularly interesting:

1) On 19 November at 0700 GMT the Froude number at Satolas and Valence are too large to be compatible with our scheme. The results for the rest of the day, however, show that the flow structure changes and later shows much better agreement with our assumptions.

2) On 19 November at 1200 GMT the Froude number at Orange is very large while the inversion layer base remains low. This leads to the conclusion that the jump has occurred downstream of Orange, as confirmed by the data taken at Nîmes.

It is our feeling that the theory can describe fairly accurately the flow up to the point where the jump occurs. Downstream of it, however, it seems that the theory is not satisfactory, partly because of the orography of the southern part of the Rhône Valley and also because of some difficulties in quantifying the downstream conditions.

The occurrence of a jump indeed divides the flow into two separate regimes. The problem is then one of dealing with the sharp or progressive nature of the transition between these regimes. The kinetic energy of the mean flow which is lost at the jump is either transformed into turbulent kinetic energy or radiated away by a stationary wave system located downstream of the jump (Ball, 1956). The aircraft cross sections shown in Fig. 3 are consistent with such a mechanism, specially because of the strong contrast between the northern and southern parts of the valley, of the undulatory nature of the southern flow, and of the sharp change in regimes near Montelimar.

As a last element to confirm this interpretative scheme, we tried to measure the turbulent fluxes when possible since one of the two aircraft was equipped with gust probes. The inhomogeneity of the flow, however, is unfortunately such that estimated momentum fluxes are not reliable.

6. Conclusion

The analysis of the "Campagne Vallée du Rhône 77" has brought some new elements for the phenomenological study of certain violent winds in the Rhône Valley, particularly by providing a continuous record of profiles of the main meteorological variables along the north–south axis of the valley.

A detailed analysis of the days with a north wind shows very coherent measurements, with a strong north–south contrast and indicates that the flow is stationary. These features led us to interpret the observed characteristics with the aid of a phenomenological model of airflow beneath an inversion layer in a narrow valley. The violent winds in the Rhône Valley can then be described as follows:

- In the north of the valley, between Satolas and Valence, the winds are relatively weak and the height of the inversion layer remains roughly constant.
- In the region of the constriction, near Valence, the wind speed accelerates suddenly and the height of the inversion layer is greatly reduced.
- In the southern part of the valley, between Montelimar and Nîmes, a jump occurs downstream of which the wind is slowed down by a significant amount and the slowly varying height of the inversion layer is higher than that upstream of the jump but lower at the entry of the valley.

The experimental measurements confirm qualitatively the model results and indicate that the jump occurs most of the time between Montelimar and Orange. As far as the violent winds of the Rhône Valley are concerned, our analysis indicates that:

- The shape of the Rhône Valley, narrowing to half-width in the region of Valence and the presence of an inversion layer would lead to the appearance of violent winds in the south of the valley, with a speed five times greater at Orange than at Satolas, if no jump was going to take place.
- The further downwind the jump, the stronger the wind in the south of the valley, in spite of what
could then be expected from the widening of the valley.

- The flow upstream of the jump is governed only by upwind conditions; the location of the jump in the south of the valley can be determined from the height of the inversion layer at the valley exit. The lower the ratio between the height of the inversion layer at the exit and the corresponding height at the entry of the valley, the further downwind the jump and consequently the stronger the winds in the south. It may thus be postulated that synoptic situations characterized by subsidence over the Mediterranean will be very favorable for the creation of violent winds on the coast.

A climatological study would be essential to show if situations with temperature inversions are frequent enough to confirm the validity of the theory. Such a study would be complicated by the fact that the synoptic network is too coarse to provide good criteria for the selection of strong wind situations, even though the present phenomenological description shows that the flow can take various aspects without altering the physics of the problem.

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