Vertical Wind Component Estimates up to 1.2 km above Ground

LAURENCE J. RIDER AND MANUEL ARMENDARIZ
Atmospheric Sciences Lab., U. S. Army Electronics Command, White Sands Missile Range, N. M.

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ABSTRACT
Vertical wind components were computed up to 1.2 km from 37 wintertime and 10 summertime balloon observations between 0900 and 1200 local time utilizing the accurate and high-resolution cineheodolite/Jimsphere system. The mean ascent rate of the Jimsphere was computed from all observations taken on a particular day. The ascent rate was found to be 5.16 m sec\(^{-1}\) for the winter and 5.10 m sec\(^{-1}\) for the summer months. The individual variations of a given observation from the mean ascent rate were assumed to be the vertical component. Variations in balloon ascent caused by variation in drag, anomalous variation in atmospheric density, balloon response to the wind, and aerodynamically induced motions are discussed. Vertical wind components ranged from 10–25 cm sec\(^{-1}\) in a stable atmosphere and 35–100 cm sec\(^{-1}\) under unstable conditions depending on wind speed.

1. Introduction

Considerable attention has been given to the vertical component of atmospheric turbulence in the very lowest few meters of the atmosphere, but far too little study has been made of the fluctuations of the vertical wind component above meteorological tower levels. There are two reasons for this; first, the parameters are most accessible to measurements at tower heights; and second, a convenient organization of the data has developed gradually from the standpoint that the turbulent properties are characterized by the vertical fluxes of momentum and heat, both of which are assumed constant within the lowest few meters of the atmosphere. As a result, there are available useful, although empirical, generalizations (Lumley and Panofsky, 1964) from which certain statistical properties of turbulence are prescribable, given a knowledge of the terrain and general weather conditions. At greater heights, above ~100 m, conditions are very different, and the acquisition of acceptable quality data is much more difficult. Moreover, the simplifications of constant stress and flux are not applicable. The understanding of the mechanics of turbulent motion presents a much more difficult task above the surface boundary layer. Nevertheless, the deeper layer is of considerable importance in meteorological problems, especially in the sense that its properties determine the median-range spread of atmospheric contaminants.

Until recently, there have been only three main sources of direct measurement of vertical wind components at heights above tower levels. The three techniques have primarily involved measurements from vanes on tethered balloons (Smith, 1961), the use of a slow aircraft for a platform (Bunker, 1956), and a sailplane utilized in the measurement of turbulence (MacCready, 1962). Recently, DeMandel and Krivo (1968) investigated the capability of the FPS-16 radar/Jimsphere system for direct measurement of vertical air motions up to an altitude of 15 km. They concluded that a measure of the vertical wind component may be derived from Jimsphere (2 m diameter superpressure balloon of constant volume) ascent-rate data despite the high noise level in the original data, and despite the difficulty in distinguishing between balloon response to buoyancy and to vertical air motions.

Most of the data in this study were obtained utilizing 2 m diameter Jimspheres, except on 10 July 1967 when 1 m diameter Jimspheres were used. Some important techniques used in this study that differ from those used by DeMandel and Krivo are: 1) the tracking of the Jimsphere balloon was accomplished by accurate and high-resolving cineheodolite cameras (three or four) instead of the FPS-16 radar, which, according to Scoggins and Armendariz (1969), contains many sources of errors; and 2) the data extended to an altitude of only 1.2 km as compared to 15 km in the DeMandel and Krivo study. In addition, there were some differences in the treatment of the data, but the objectives were basically the same. For example, the data in this paper were smoothed by a least-squares 11-point moving-arc filter, whereas DeMandel and Krivo fitted a first-degree polynomial by the least-squares method as done by Scoggins (1962). Smoothing techniques virtually eliminated high-frequency (>0.22 cps), aerodynamically induced balloon motions from the data as well as systems error. The average unsmoothed system position error was 15.8 cm with a maximum error near 30.5 cm for the data used in this paper. It will be shown that variations in the ascent rate of a constant volume balloon such as the Jimsphere due to density anomalies encountered during a particular flight up to 2 km are insignificant.

One of the primary applications of the results of this kind of vertical wind measurement is in the area of particle diffusion, particularly to mesoscale motions under conditions of a thermally stratified atmosphere. The existence or lack of a vertical current through a surface temperature inversion to stronger horizontal
winds above can determine the rate of diffusion of contaminants released near the surface.

2. Collection and treatment of data

The data in this study were obtained from a number of Jimsphere balloons which were tracked by three or four cinetheodolite cameras on two wintertime days, 29 November 1966 and 12 December 1966, and two summertime days, 10 July 1967 and 5 August 1968.

Cinetheodolite position measurements in spherical coordinates were recorded at 1-sec intervals and were smoothed by an 11-point, moving-arc, second-degree polynomial. The smoothed data were then converted to Cartesian coordinates, and the x, y, z component velocities of the balloon movement were computed (see Appendix).

The mean z component of the balloon movement was calculated for each run (made at 6-min intervals), and an average of the means for all runs during a period of operation was assumed to be the mean ascent rate of the balloon for that particular period. For example, on 29 November 1966, the mean ascent rate was 5.16 m sec\(^{-1}\) for all of the 2 m Jimspheres released in a 2\(\frac{1}{2}\)-hr period. An identical mean ascent rate was obtained for all ascents during a 2-hr period on 12 December 1966. The same procedure was followed with a series of 1 m Jimspheres released on 10 July 1967, and the mean ascent rate was 4.23 m sec\(^{-1}\). On 5 August 1968, 2 m Jimspheres were used, and the mean ascent rate was found to be 5.10 m sec\(^{-1}\).

The mean ascent rate for all Jimspheres runs conducted during an \(\sim 2\)-hr period on a specific day (such as 29 November 1966, 5.16 m sec\(^{-1}\)) was then subtracted from individual ascent rate profiles at each second data point and the differences assumed to be largely due to the effects of vertical air motions.

These differences, which will be called vertical wind \(w\), are considered to be downward if negative (less than the mean for the series on that particular day) and upward if positive (greater than the mean). These differences were algebraically averaged through 100 m layers from the surface to 1200 m, yielding a net mean upward or downward wind component for each 100 m layer in the algebraic averaged results, or in the case where the sign (up or down) was disregarded, an absolute magnitude of the vertical components was obtained for each 100 m layer.

3. Discussion and results

Before one assumes that the variations in ascent rate in individual runs from the mean ascent rate for numerous runs taken during a 2-hr period are due to vertical winds alone, one should consider known factors which may influence balloon motion. These include: 1) aerodynamically induced balloon motions, 2) balloon response to vertical shear of the horizontal wind, 3) icing and condensation, 4) variations in size and/or mass of the balloon, 5) anomalous variation in atmospheric density, and 6) variation in the drag coefficient.

The Jimsphere experiences aerodynamically induced oscillations having a wavelength of \(\sim 22\ m\) (cf. Rogers and Cannitz, 1965; Armendariz and Rachele, 1967; Rider and Armendariz, 1968). Most of these oscillations were removed by the smoothing technique used in this
study (see Appendix). Extreme wind shears which are necessary to produce a significant change in the vertical motion of a Jimsphere must occur over a very thin layer and therefore their effect would be removed by the smoothing performed on the data (DeMandell and Krivo, 1968).

The data used in this study were obtained on days when the moisture content was far below saturation at the levels considered, and there was no precipitation in the vicinity at the time of balloon ascent.

The Jimsphere balloon is a constant volume balloon, and it can be safely assumed within the limits of the data used, up to 1200 m, that there was no significant change in the size or mass of the balloon and its helium content. The response characteristics of the Jimsphere have been determined by Eckstron (1965) in terms of wind gradient at several altitudes. For this study, the Jimsphere was capable of detecting a wind shear of 0.1 sec⁻¹ which is the strongest wind gradient expected with a wind response error of 10 cm sec⁻¹.

DeMandell and Krivo developed an expression for balloon ascent $V_s$ by considering the forces on a balloon rising through a quiescent atmosphere. The resultant expression (cgs units) can be written as

$$C_D V_s^2 = 2.26 \times 10^8 \left( \frac{63.1}{25.5 + \frac{T}{T_h}} \right) + 3.61 \times 10^8 \frac{t}{T_h},$$  \hspace{1cm} (1)

where $C_D$ is the coefficient of drag, $V_s$ the balloon ascent rate, $\rho$ density, $T_h$ the internal temperature of the helium, $T$ the ambient temperature, and $t$ the temperature deviation, i.e., $T_h = T + t$. [Since in this study $T_h$ varied from $\sim 250$ to near $300 K$, $\rho$ varied from $\sim 1.07 \times 10^{-3}$ to $0.92 \times 10^{-3}$ gm cm⁻³, and $t$ was generally less than 10K, the above expression was simplified to

$$C_D V_s^2 = \frac{1}{\rho} - (25.5),$$  \hspace{1cm} (2)

because the two terms dropped from (1) are approximately two orders of magnitude smaller than the others, as found by DeMandell and Krivo.] The graphical presentation (Fig. 1) of $C_D$ vs vertical wind speed for the limiting values of density in this study shows that for a given $C_D$, $V_s$ can change approximately 5 cm sec⁻¹ due to the extreme change in density shown. However, it is improbable that such an extreme change of density would be encountered in the general atmosphere over small distances ($< 50$ m), and one must conclude that anomalous variations in atmospheric density, within the time limits of this study, will cause only minor changes in ascent rate which can be safely disregarded.

If one considers the effect of wind shears or change in drag, DeMandell and Krivo found that, for the Jimsphere, extreme wind shears would change the vertical velocity of the balloon by no more than 1.5 cm sec⁻¹. Moreover, they stated that a wind shear that would produce such a change in $V_s$ would extend over a small
vertical distance and that filtering or smoothing would remove this effect. The coefficient of drag was found to be almost constant, approximately 0.760 at White Sands through the first 1.2 km above the surface. This value of $C_D$ compares with that of DeMandel and Krivo who found $C_D$ to be 0.730 for observations at Cape Kennedy.

Two examples of the magnitude and variation of the vertical component of the balloon motion are shown in Figs. 2a and 2b. Variation of the ascent rate of the balloon released at 0945 MST 29 November 1966 is shown in Fig. 2a where the vertical solid line represents the mean ascent rate (5.16 m sec$^{-1}$) for all the 2 m balloons released in a 2½-hr period. Fig. 2b shows the variation of the ascent rate of the balloon released at 1115 MST 5 August 1968, where the mean ascent rate is represented by the vertical solid line (5.10 m sec$^{-1}$).

Figs. 3-5 show the variation of $\bar{w}$ (by 100 m layers) with height during periods of different degrees of thermal stability. At the same time, reference should be made to Fig. 7 in which temperature profiles are plot-
ted from radiosonde observations taken in the same area.

Fig. 3 shows the first two and the last two releases of a series of 20 Jimsphere runs spaced 6 min apart on 29 November 1966. As the instability increased in the lower layers, the magnitude of the mean vertical winds increased from \( \sim 10-25 \text{ cm sec}^{-1} \) to 55-85 cm sec\(^{-1}\). Similarly, in Fig. 4 there is also a marked increase in vertical winds with an increase in instability during another series of 20 cinetheodolite/Jimsphere runs on 12 December 1966. Continuing this analogy, Fig. 5 presents the vertical variation of the vertical wind for two cinetheodolite/Jimsphere runs an hour apart on 10 July 1967, and Fig. 6 two runs an hour apart on 5 August 1968. Vertical wind components are stronger, and updraft and downdrafts are sustained through deeper layers with decreasing stability. Further variation of vertical wind with height and stability can be seen in
Fig. 7. Temperature profiles for 0800 and 1200 MST on 29 November 1966 and 12 December 1966, and for 0800 MST on 10 July 1967 and 5 August 1968.

Fig. 8, where the absolute means $|\bar{w}|$ are plotted by 100 m layers for all the soundings made on a particular day.

For study of the behavior of the standard deviation $\sigma_w$ of the vertical wind above the surface boundary layer, there are primarily the measurements by Smith (1961) from vanes mounted on tethered balloons and by Bunker (1956) by means of a slow aircraft. Fig. 9

Fig. 8. Absolute means of the computed vertical wind for particular days.

Fig. 9. The ratio of $\sigma_w$ to $\bar{w}$ as a function of $\bar{w}$ and stability between 500 and 5000 feet at Cardington (after Smith, 1961).
reproduces Smith's observed relation between the standard deviation of vertical angle, wind speed, and stability. The variation of \( \sigma_w \) (with stability) was objectively confirmed by associating with each point the stability deduced from observations of temperature gradient in the vertical, made as near as possible to the time of record. This study agrees in a general way with Fig. 9, as can be seen in Fig. 10, which was plotted from cinetheodolite/Jimsphere data. The difference in the methods of arriving at a stability classification by Smith and in this paper must be pointed out. Smith obtained the overall lapse rate from the ground up to the vane, whereas in this study, the lapse rate was considered only for individual 100 m layers where \( \omega, \sigma_w, \) and the horizontal wind \( u \) were computed. In addition, Smith's measurements of \( \omega, \sigma_w \), and \( u \) are Eulerian measurements. The data in this study cannot be strictly Eulerian or Lagrangian, but perhaps more nearly Lagrangian than Eulerian. Much of the data were collected in the winter with stable lapse rates and relatively light winds. This gave several points with the ratio \( \sigma_w/\bar{u} < 0.10 \) and wind speeds \(< 8 \text{ m sec}^{-1} \). However, there were several situations with stable lapse rates and horizontal winds up to nearly \( 17 \text{ m sec}^{-1} \).

4. Conclusions

It appears that vertical wind components can be estimated within the first few hundred meters above the surface by utilizing the cinetheodolite/Jimsphere system with suitable smoothing techniques. The magnitude of these vertical wind components is apparently controlled to a large degree by the thermal stability of the atmosphere. In unstable air the vertical winds are much stronger (\( 55-100 \text{ cm sec}^{-1} \)) and the sign (up or down) persists through much deeper layers than when the air is stable. Under stable conditions, vertical wind components are of the order of \( 10-25 \text{ cm sec}^{-1} \). This agrees well with accepted theory since convective turbulence has a much longer wavelength than mechanical turbulence. The variability of the vertical wind is a function of stability to some degree, as well as of the horizontal wind speed.

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APPENDIX

Frequency Response for the 11-Point, Second-Degree Polynomial Smoothing

This smoothing is accomplished by fitting a second-degree polynomial to 11 consecutive points by least squares and evaluating the polynomial at the midpoint. This process is continued by “slipping” one point until all the data are smoothed (except the first and last five points).
The normal equations for the least-squares fit are as follows:

\[ \Sigma x = a \Sigma t^2 + b \Sigma t + nc, \]

\[ \Sigma tx = a \Sigma t^3 + b \Sigma t^2 + c \Sigma t, \]

\[ \Sigma t^2 x = a \Sigma t^4 + b \Sigma t^3 + c \Sigma t^2. \]

The computations are considerably simplified by continually "slipping" the \( t \) axis so that the midpoint of the 11 points is defined to be \( t=0 \). Observe that \( \Sigma t^n = 0 \) for \( n \) odd. Also the parabola

\[ X_n = a t^2 + b t + c \]

degenerates to \( X_n = c \) since the smoothed value is at the midpoint \( (t=0) \).

The normal equations become

\[ \Sigma x = a \Sigma t^2 + nc, \]

\[ \Sigma tx = b \Sigma t^3, \]

\[ \Sigma t^2 x = a \Sigma t^4 + c \Sigma t^2. \]

Hence

\[ c = \frac{\Sigma t^2 \Sigma t^2 x - \Sigma t^4 \Sigma x}{(\Sigma t^2)^2 - n \Sigma t^4}, \]

which can be rewritten as

\[ c = \frac{\sum (t^2 \Sigma t^2 - \Sigma t^4) X}{(\Sigma t^2)^2 - n \Sigma t^4}. \]

In terms of filter weights, since \( \Delta t = 1 \), this becomes

\[ X_n = \frac{\sum \left( t^2 \Sigma t^2 - \Sigma t^4 \right) X_i}{(\Sigma t^2)^2 - n \Sigma t^4} = \sum W_i X_i, \]

where \( X_n \) is the smoothed value and each of the summations is from \( i = 1 \) to 5. Performing the arithmetic gives (observe \( W_{-i} = W_i \)):

\[ W_0 = 1958/9438, \quad W_1 = 1848/9438, \quad W_2 = 1518/9438 \]

\[ W_3 = 968/9438, \quad W_4 = 198/9438, \quad W_5 = 792/9438 \]

Since \( \Delta t = 1 \), the filter response is

\[ R(f) = W_0 + 2 \sum_{i=1}^{5} W_i \cos 2\pi f \]

\[ = \frac{1}{9438} \{1958 + 2[1848 \cos 2\pi f + 1518 \cos 4\pi f + 968 \cos (6\pi f) + 198 \cos 8\pi f - 792 \cos 10\pi f] \}. \]

REFERENCES


