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Analysis of the diurnal oscillation of surface geostrophic wind over Western Europe

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ABSTRACT

An analysis of surface meteorological data, recorded by the West-European synoptic network in June and July 1976, reveals the existence of a diurnal oscillation of surface geostrophic wind in this part of Europe. The oscillation is similar to that observed in the Great Plains of the United States by several investigators.

The analysis delineates physical factors involved in this diurnal oscillation of geostrophic wind. As found by others, the oscillation is induced partly by the diurnal temperature cycle over sloping terrain. Our analysis suggests that a diurnal variation of the horizontal gradient of the temperature field in the atmospheric boundary layer is another major factor contributing to the observed oscillation of geostrophic wind.

1. Introduction

Diurnal wind oscillation in the atmospheric boundary layer has long been investigated, especially over the Great Plains of the United States where the boundary layer wind frequently exhibits a pronounced nocturnal maximum, called nocturnal low-level jet (Means, 1952). Frequent night-time storms in this region can also be explained by the formation of a nocturnal convergence field associated with this low-level jet (Sangster, 1967; Bonner, 1966; Wallace, 1975). Although a number of different theories have been proposed to explain the occurrence of both phenomena and their apparent association (Buajitti and Blackadar, 1957; Wexler, 1961; Holton, 1967), it now seems firmly established that the gentle sloping terrain, from the Rocky Mountains to the Mississippi River, is one of major physical factors involved in these phenomena (Bonner, 1968; Lettau, 1967; Bonner and Paegle, 1970). In particular, the theory proposed by Holton (1967) emphasized the contribution of air drainage along the sloping terrain to the diurnal evolution of wind profile in the boundary layer; the drainage is induced by the diurnal cycle of temperature over the sloping terrain.

This mechanism involves a diurnal oscillation of geostrophic wind in the atmospheric boundary layer, which is easy to verify by synoptic network data analysis. By employing surface and upper air meteorological data collected over the Great Plains, Hoecker (1965), Sangster (1967) and Bonner and Paegle (1970) clearly confirmed the existence of such an oscillation. Lettau (1967), Bonner and Paegle (1970), and Paegle and Rasch (1973) showed that, when compared with experimental results, the diurnal evolution of boundary layer wind which was numerically simulated could be improved in its amplitude and phase by taking into account this diurnal dynamic forcing function. Furthermore, it was shown that the dynamic forcing significantly affects the turbulent momentum transfer, both in the planetary boundary layer (Hoxit, 1974) and in the surface boundary layer (Caughey *et al.*, 1979).

Isaka and Pinty (1981) reported the occurrence of a diurnal oscillation of geostrophic wind over the western part of Europe similar to that observed over the Great Plains. It appeared, however, that this diurnal oscillation could not be explained alone by the diurnal temperature cycle over the large-scale topographic slope existing in this region. The purpose of the present paper is to identify other physical factors involved in this phenomenon.

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2. Data and methods

The data we used in this analysis are derived from West-European meteorological network stations; namely surface data taken every 3 h and semi-diurnal upper air soundings. After checking 3 months' data (from May to July 1976), we selected 1 month's data (from 15 June to 15 July). Missing or obviously erroneous data were replaced by values linearly interpolated from adjacent data. Less than 9% of the total amount were missing or erroneous. It should be noted that this period was in the middle of the severe drought which affected the whole of western Europe in 1976.

The mean value and the horizontal gradient of a meteorological variable at a given geographical site were computed by adjusting a quadratic function to its values obtained at neighbouring stations. In this adjustment, the weighting function defined by Cressmann (1959) was used with a value of 200 km for the radius of influence. This value is a compromise between the need to have sufficient number of stations for the adjustment and the wish to use a radius as small as possible. This may cause a relatively strong smoothing of the fields, especially near the coast where relatively large wind and temperature gradients can be expected. This smoothing, however, may not significantly alter the main features of the phenomenon analysed in the following. The surface geostrophic wind was computed at 7 geographical sites: Kleine Brogel (51°10′N, 5°28′E), St. Quentin (49°49′N, 3°12′E), Saarbrücken (49°13′N, 7°07′E), Pithiviers (48°19′N, 2°10′E), Angers (47°30′N, 0°34′W), and Angoulême (45°40′N, 0°13′E).

Spectra presented in this paper were calculated from a time series of the difference:

$$\delta X = X(t) - X(t + 3 \text{ hours}) \tag{1}$$

in order to filter out low-frequency components. Since $S(\delta X) \propto S(X)f^2$, where $S(\delta X)$, $S(X)$ and f designate the spectral densities of the δX- and X-series, and the frequency respectively, from here on we will consider the δX spectrum as the X spectrum, represented with $S(X)f^2$ as the ordinates. Because of the small total number of data points in the time series, spectra were estimated using the Blackmann–Tukey correlation method. A program given by Jenkins and Watts (1969) was used with the Parzen window to avoid negative spectral densities. The number of degrees of freedom and the width of the frequency band were 19 and 0.21 cycles/day respectively for all the computed spectra.

3. Description of the diurnal oscillation of surface geostrophic wind

The surface geostrophic wind can be expressed in the following form (Sangster, 1967; Bonner and Paegle, 1970):

$$U_g = -\frac{g}{f}(1 + S^*)\frac{\partial D}{\partial y} + \frac{g}{f}S^*\frac{\partial Z_s}{\partial y}$$

$$V_g = \frac{g}{f}(1 + S^*)\frac{\partial D}{\partial x} - \frac{g}{f}S^*\frac{\partial Z_s}{\delta x} \tag{2}$$

where

$$D = Z_s - Z_p, \qquad S^* = \frac{T_s - T_{sp}}{T_{sp}}.$$

g, Z_s and T_s are the apparent gravitational acceleration, elevation of terrain and virtual temperature at terrain surface respectively. Z_p and T_{sp} represent the altitude and temperature respectively in the standard atmosphere corresponding to the surface pressure, and U_g and V_g are the east–west and north–south components of the surface geostrophic wind.

Figs. 1a,b represent the δU_g and δV_g spectra of the surface geostrophic wind estimated at the 7 sites. At every site, the δU_g spectrum exhibits a relative maximum near 0.3 cycle/day, corresponding to the variation of synoptic pressure field, and another well-defined maximum at the diurnal frequency (1 cycle/day). The most pronounced δU_g peak is found at Kleine Brogel. At every site, the δV_g spectrum shows a well-defined diurnal oscillation, without any significant spectral density near 0.3 cycle/day. The largest diurnal oscillation of the V_g component occurs at Angoulême, while the smallest oscillation is observed at Saarbrücken. Figs. 1a,b also show very small peaks at the frequencies corresponding to 2 to 3 cycles/day. Since these spectra are computed from the time-series of the difference δX, the effective contribution of these higher harmonics to the diurnal oscillation may be estimated by dividing the corresponding spectral densities by (frequency)². This means that the oscillation of surface geostrophic wind is basically diurnal, without any significant contribution from the higher harmonics. This

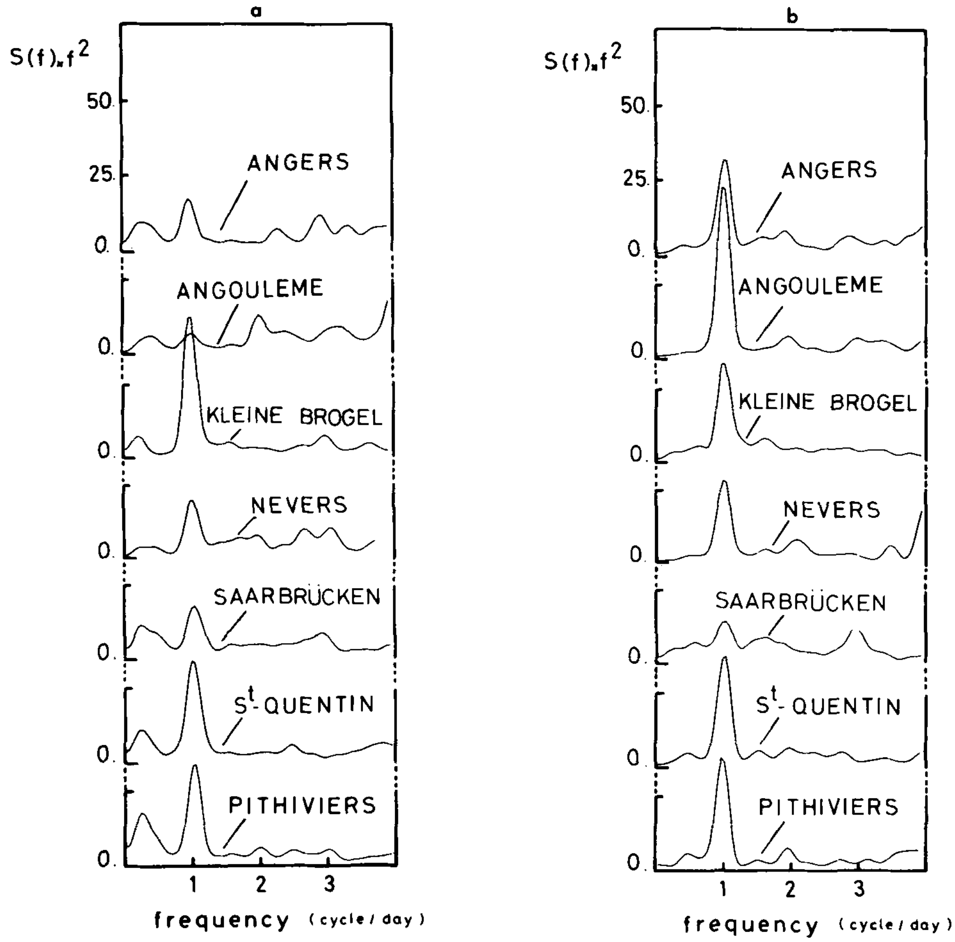

Fig. 1. The geographical variation of the δU_g (a) and δV_g (b) spectra, computed from the time series of the surface geostrophic wind. $S(f)$ represents the spectral density of the U_g and V_g spectra. The ordinates are given in arbitrary units.

absence of the semi-diurnal oscillation in the δU_g and δV_g spectra suggests that the semi-diurnal oscillation of the surface pressure, which is clearly defined in the δP_s spectrum and is mainly due to the semi-diurnal solar tide (Pinty, 1980), is relatively uniform in its amplitude and phase from one site to another.

4. Physical factors involved in the diurnal oscillation of the geostrophic wind

4.1. *Effect of a large-scale terrain slope*

As previously mentioned, one of the major physical factors involved in the diurnal oscillation of geostrophic wind is the diurnal cycle of temperature over a sloping terrain (Lettau, 1967; Sangster, 1967; Bonner and Paegle, 1970). According to Bonner and Paegle, the *direct effect* of the sloping terrain is represented by the second term on the right-hand side of eq. (2). This term is a product of the terrain slope with the specific diurnal variation of surface temperature.

After elimination of the long-term tendency by subtracting the 24 h running-averaged values from the original data, the mean diurnal variations of each term of eq. (2) were computed for the period from June 28 to July 8 at all sites. Fig. 2 shows the geographical variation of the mean diurnal oscil-

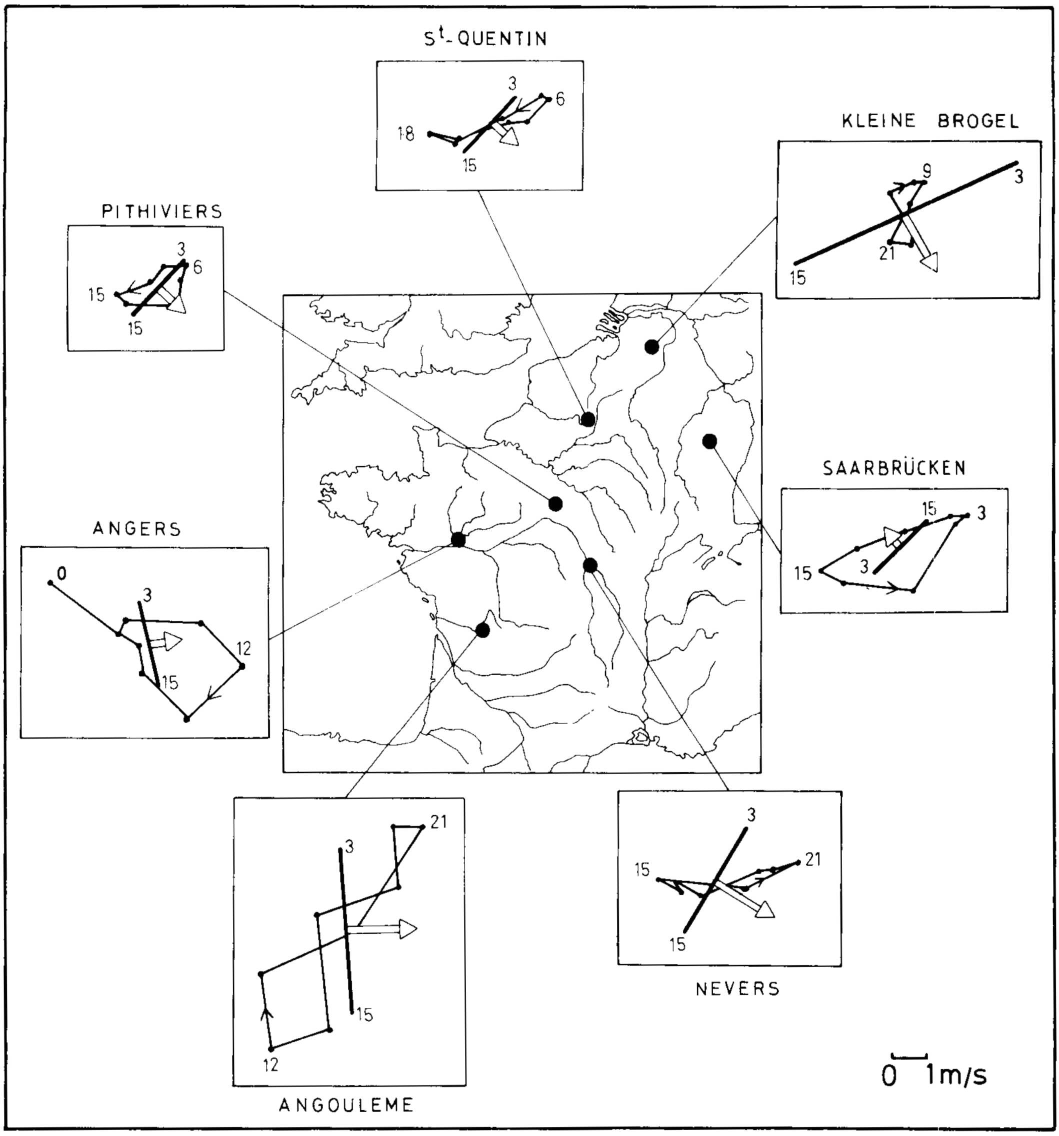

Fig. 2. Hodograph of the mean diurnal variation of the first term (thin line) and second term (thick line) of eq. (2) (from June 28 to July 8). On each hodograph, the numbers indicate GMT.

lations of the first and second terms. We will examine at first the variation of the second term which represents the direct topographical effect. The geographical variation of the first term will be discussed in the subsequent paragraph.

In Fig. 2, the terrain slope estimated at each site mainly reflects the gradual rising of terrain from the English Channel at the Bay of Biscay to the Alps and the French Massif Central, while at Saarbrücken the reversed slope results from the proximity of the Ardennes located north-west of this site. The oscillation occurs perpendicular to the terrain slope, as indicated by the hollow arrow. The extreme values of the oscillation take place at the times of minimum and maximum surface temperature, namely at 03 and 15 GMT. The wind vector corresponding to the maximum temperature is directed so that the higher terrain is on its left. The amplitude of the geostrophic oscillation depends both on the terrain slope and the amplitude of surface temperature variation. Thus, the difference between the amplitudes at Angoulême (2.5 m s^{-1}) and Kleine Brogel (3.5 m s^{-1}) results as much from the difference in terrain slope ($9/10^5$ at Angoulême and $11/10^5$ at Kleine Brogel), as from that in amplitude of the diurnal temperature cycle (12 °C at Angoulême and 16 °C at Kleine Brogel). These amplitudes of diurnal oscillation of geostrophic wind, due to the sloping terrain, are quite comparable to those obtained by Bonner and Paegle (1970) over the Great Plains. These authors gave a value of about 3 m s^{-1} as the direct contribution of the topographic effect.

Fig. 2, however, shows that the contribution of the first term to the diurnal oscillation of surface geostrophic wind is generally as large as that of the second term. This is quite different from Bonner and Paegle's results, because in their case the amplitude of the first term did not exceed a value of 1.5 m s^{-1}.

4.2. *Effect of the horizontally non-uniform diurnal temperature cycle*

The spectral analysis of the first term of eq. (2) confirmed the diurnal characteristic of its short-period oscillation, without any significant contribution of the higher order harmonics (Isaka and Pinty, 1981). Fig. 2 shows that the extreme values of this oscillation occur slightly out of phase from those of the surface temperature (03 and 15 GMT). The corresponding times vary from one site to another, i.e. 12 and 21 GMT at Angoulême, 06 and 18 GMT at St. Quentin, 00 and 12 GMT at Angers, etc. The amplitude of the oscillation also varies from site to site without any apparent relationship to the amplitude of the second-term oscillation. Thus the amplitude of the first-term oscillation is smallest at Kleine Brogel (1 m s^{-1}) and largest at Angoulême (3.5 m s^{-1}), while the second-term oscillation is found to be of the same order at both these sites. The direction of the oscillation is significantly shifted compared to that of the second term. The shift is about 60° at Nevers and 180° at Saarbrücken, while both the first and second terms oscillate practically in the same directions at Pithiviers and St. Quentin. These results suggest that physical factors other than the temperature cycle over the sloping terrain are involved in the diurnal oscillation of surface geostrophic wind, and that their contributions are as large as that of the topographical effect.

By considering the diurnal temperature perturbations, air density and atmospheric pressure around their daily average values and by neglecting the contribution of second order terms, we can express the first term (U_{g1}) of eq. (2) in the following form:

$$U_{g1} = \bar{U}_{g1} - \frac{\rho'_s}{\bar{\rho}_s}\bar{U}_{g1} - C\frac{\partial Z_s}{\partial y} - \frac{1}{f\bar{\rho}_s}\frac{\partial P'_s}{\partial y} \tag{3}$$

where

$$C = \frac{g}{f}\frac{\bar{T}_s}{\bar{T}_{sp}}\left(\frac{g}{R\gamma} - 1\right)\frac{T'_{sp}}{\bar{T}_{sp}}.$$

Primed and over-barred variables represent diurnal perturbations and daily averages, respectively. R and γ designate the gas constant of the air and the lapse rate in the standard atmosphere. A similar expression can be written for the V-component. Since the first term on the right-hand side of eq. (3) represents the daily average of the U_{g1} term, the diurnal variation should result from its second to fourth terms.

The second term of eq. (3) represents the influence of the diurnal air density variation on the U_{g1} term. The air density variation results essentially from the diurnal air temperature variation. Since the amplitude and direction of this term directly depend on the daily average of the U_{g1} term, its contribution may introduce a shift in the direction of diurnal oscillation. In the present case where the average U_{g1} wind and the peak to peak amplitude of T'_s were at most of the order of 10 m s^{-1} and 18° C respectively, the amplitude of this second term remains less than 0.3 m s^{-1}, one-fifth to one-tenth of the amplitude of the U_{g1} oscillation.

The third term of eq. (3) depends both on the terrain slope and the specific temperature variation (T'_{sp}) in the standard atmosphere, which arises from the diurnal perturbation of surface pressure. In the present case, the pressure perturbation was of the order 2 mb. This corresponds to a temperature fluctuation of 0.1 °C, about one-hundredth of the surface temperature variation (T'_s). Since the factor ($\bar{T}_s/\bar{T}_{sp}$) is of order unity, the contribution of the third term to the diurnal oscillation of geostrophic wind may be neglected. Our findings suggest that the diurnal oscillation of the U_{g1} term must arise from the fourth term of eq. (3), which depends on the horizontal gradient of the surface pressure perturbation (P_s').

Assuming that the pressure and density perturbation are in hydrostatic equilibrium, we can write:

$$\frac{\partial P_s'}{\partial y} = -g \frac{\partial}{\partial y} \int_{Z_s}^{\infty} \rho'(x, y, z)\, dz \tag{4}$$

where P_s' and ρ' represent the surface pressure perturbation and air density perturbation, respectively.

Since $P'/\bar{P} < T'/\bar{T}$, the density perturbation due to diabatic heating can be written as

$$P' \simeq -\frac{\bar{P}}{\bar{T}} T'. \tag{5}$$

In order to pursue further analysis in a mathematically simple, but nevertheless physically relevant way, we assume a diurnal heating function which damps the temperature perturbation exponentially with height (Holton, 1967; Paegle, 1978):

$$T' = A \exp\left\{\frac{-(z - z_s)}{H}\right\} \exp(i\omega t) \tag{6}$$

with $\omega = 2\pi/24$ h. A and H are functions of horizontal coordinates. A represents the amplitude of the diurnal surface temperature cycle and H the characteristic height of the heating function. In his paper, Paegle (1978) adopted a value of 300 m for H. Wallace and Patton (1970), showed that the height H in the High Plateau is about twice as large as that in the Mississippi Valley.

By substituting eqs. (5) and (6) into eq. (4), we obtain:

$$\left(\frac{\partial P_s'}{\partial y}\right)_Z = g \frac{\partial}{\partial y} \int_{Z_s}^{\infty} \frac{\bar{\rho}}{\bar{T}} A \exp\left\{\frac{-(z - Z_s)}{H}\right\} \exp(i\omega t)\, dz. \tag{7}$$

In order to calculate this integral explicitly, we have to assume a simple mathematical function for the variation of temperature with height. If a constant lapse rate is used, the factor $(\bar{\rho}/\bar{T})$ can be expressed as:

$$\frac{\bar{\rho}}{\bar{T}} = \frac{\bar{\rho}_s}{\bar{T}_s}\left(\frac{\bar{T}_s - \gamma(z - Z_s)}{\bar{T}_s}\right)^{\gamma_a/\gamma - 2} \tag{8}$$

where γ_a and γ are the autoconvective lapse rate and the actual lapse rate, respectively. We further assume that $\gamma_a/\gamma = n$, n being an integer larger than or equal to 4, to avoid an unstable thermal stratification for the mean temperature field and to simplify the mathematical treatment. Since $\gamma < \gamma_d = 1\,°\mathrm{C}/100$ m, we may approximate the integral in eq. (7) by the expression:

$$\frac{\bar{\rho}_s}{\bar{T}_s} AH \left\{1 - \left(\frac{\bar{T}_m}{\bar{T}_s}\right)^{n-2} \exp\left(-\frac{Z_m - Z_s}{H}\right)\right\} \exp(i\omega t), \tag{9}$$

where $\bar{T}_m$ and Z_m are the temperature at the height Z_m and the height at which the diurnal diabatic heating becomes negligible, respectively. Since the difference $(Z_m - Z_s)$ may be considered to be of the same order of magnitude as the boundary layer thickness, it can be assumed to be several times larger than H. In this event, the contribution of the exponential term in the curly brackets should be less than one-tenth of the first term. Therefore, we can neglect its contribution in the following discussion without any serious error arising.

The horizontal gradient of surface pressure perturbation in eq. (4) may be written as:

$$-\frac{1}{f\bar{\rho}_s}\frac{\partial P_s'}{\partial y} = -\frac{g}{f}\frac{1}{\bar{T}_s}\exp(i\omega t) \times \left\{\frac{AH(\gamma_s - \gamma_a)}{\bar{T}_s}\frac{\partial Z_s}{\partial y} + A\frac{\partial H}{\partial y} + H\frac{\partial A}{\partial y}\right\}, \tag{10}$$

where γ_s is the lapse rate of surface temperature. In order to estimate the magnitude of these terms, we will use the following numerical values:

$$f \sim 10^{-4}\,\mathrm{s^{-1}}, \quad g \sim 9.8\,\mathrm{m\,s^{-2}}, \quad \bar{T}_s \sim 300\,\mathrm{K},$$

$$A = 8\,\mathrm{K}, \quad H = 300\,\mathrm{m}, \quad \gamma_s = 0.005\,\mathrm{K\,m^{-1}},$$

$$\gamma_a = 0.034\,\mathrm{K\,m^{-1}}, \quad \frac{\partial Z_s}{\partial y} = 10^{-3},$$

$$\frac{\partial A}{\partial y} = 10^{-5}\,\mathrm{K\,m^{-1}}, \quad \frac{\partial H}{\partial y} = 5 \cdot 10^{-4}.$$

The 3 terms in eq. (10) were estimated as 0.1 m s^{-1} (first term), 1.3 m s^{-1} (second term) and 1.0 m s^{-1} (third term). This shows that the existence of a horizontally non-uniform diurnal temperature cycle and/or boundary layer thickness contributes significantly to the diurnal oscillation of geostrophic wind, while the horizontal gradient of mean surface temperature field does not induce any such significant oscillation. The magnitude of horizontal non-uniformity used in this estimation corresponds to a difference in the amplitude of the surface temperature cycle of 1 K and in that of the characteristic height of the heating function of 50 m between 2 sites located 100 km apart. These figures are not unreasonable in the real atmospheric boundary layer.

In Fig. 3, we show the diurnal variations of the horizontal gradient of surface temperature, calculated at the 7 geographical sites, and the isolines of the peak-to-peak amplitude of the daily surface temperature cycle. In this figure, we also reproduce the diurnal variation of the geostrophic wind due to the first term of eq. (2), in order to compare it with characteristics of the diurnal thermal field.

Fig. 3 reveals the existence of the diurnal variation in the horizontal gradient of surface temperature. The amplitude of this diurnal variation varies from 0.5 K/100 km to 1.5 K/100 km. The direction along which this gradient oscillates is approximately perpendicular to the isolines of the magnitude of the daily surface temperature cycle. The diurnal surface thermal gradient is directed towards the zone of larger magnitude of diurnal temperature variation in late afternoon (15 to 18 GMT), as expected theoretically. However, it was rather difficult to establish a definite relationship between the amplitude of the diurnal surface thermal gradient and the gradient of the diurnal amplitude of the daily temperature cycle, though the order of the gradient seems numerically reasonable. At a geographical site located within a short distance of the coastline, we may suspect a contribution of the land–sea effect to the diurnal variation of the thermal gradient. This might explain the rather large magnitude of this variation observed at Angoulême. However, we do not observe any such large oscillation of the thermal gradient at Kleine Brogel, which is situated a short distance from the North Sea. Thus, if we cannot neglect the land–sea effect in the diurnal oscillation of the horizontal thermal gradient, this land–sea effect will not be the only factor to be considered.

Fig. 3 shows that at all the sites except Angers and Kleine Brogel, the diurnal oscillation of the geostrophic wind due to the first term of eq. (2) is perpendicular rather than parallel to the oscillation of the horizontal thermal gradient. At Kleine Brogel both the oscillations are not well defined and at Angers they are almost parallel to each other. Furthermore, it appears that the greater the amplitude of the diurnal variation of the surface thermal gradient, the greater the amplitude of the diurnal variation of the first term. The phase relationship between both oscillations also seems to be fairly well established. At St. Quentin, Saarbrücken, Pithiviers and Nevers, the extremes of one oscillation coincide in time with the extremes of the other.

When we analysed the influence of the horizontal non-uniformity of the diurnal thermal field, we concluded that this non-uniformity may induce a diurnal oscillation of the first term of eq. (2), whose amplitude can be of the same order of magnitude as those obtained by analysis of synoptic data. The experimental findings tend to confirm results of the theoretical analysis. The geographical variation of the characteristic height of the heating function may differ from that of the diurnal temperature cycle amplitude, though this could not be experimentally established with data from West-European upper air sounding stations. If such a difference exists, it might account for the slight deviation from the value of 90° we noticed for the angle between the direction of the diurnal oscillation of surface thermal gradient and that of the diurnal oscillation of the U_{g1} term. Furthermore, the phase of the temperature perturbation may vary with height. Paegle (1969) found a phase lag of 2 h between the surface and 500 m. This may partly explain the slight phase shift of the extremes we noticed between the two types of oscillation in Fig. 3.

5. Conclusions

The analysis of the synoptic meteorological fields over Western Europe reveals a diurnal variation of the surface geostrophic wind. The amplitude and phase of this oscillation exhibit a strong geographical dependence.

This diurnal oscillation is quite similar to those previously discovered in the Great Plains of the United States and is partly induced by the diurnal temperature cycle over sloping terrain (Hoecker, 1965; Sangster, 1967; Lettau, 1967; Bonner and

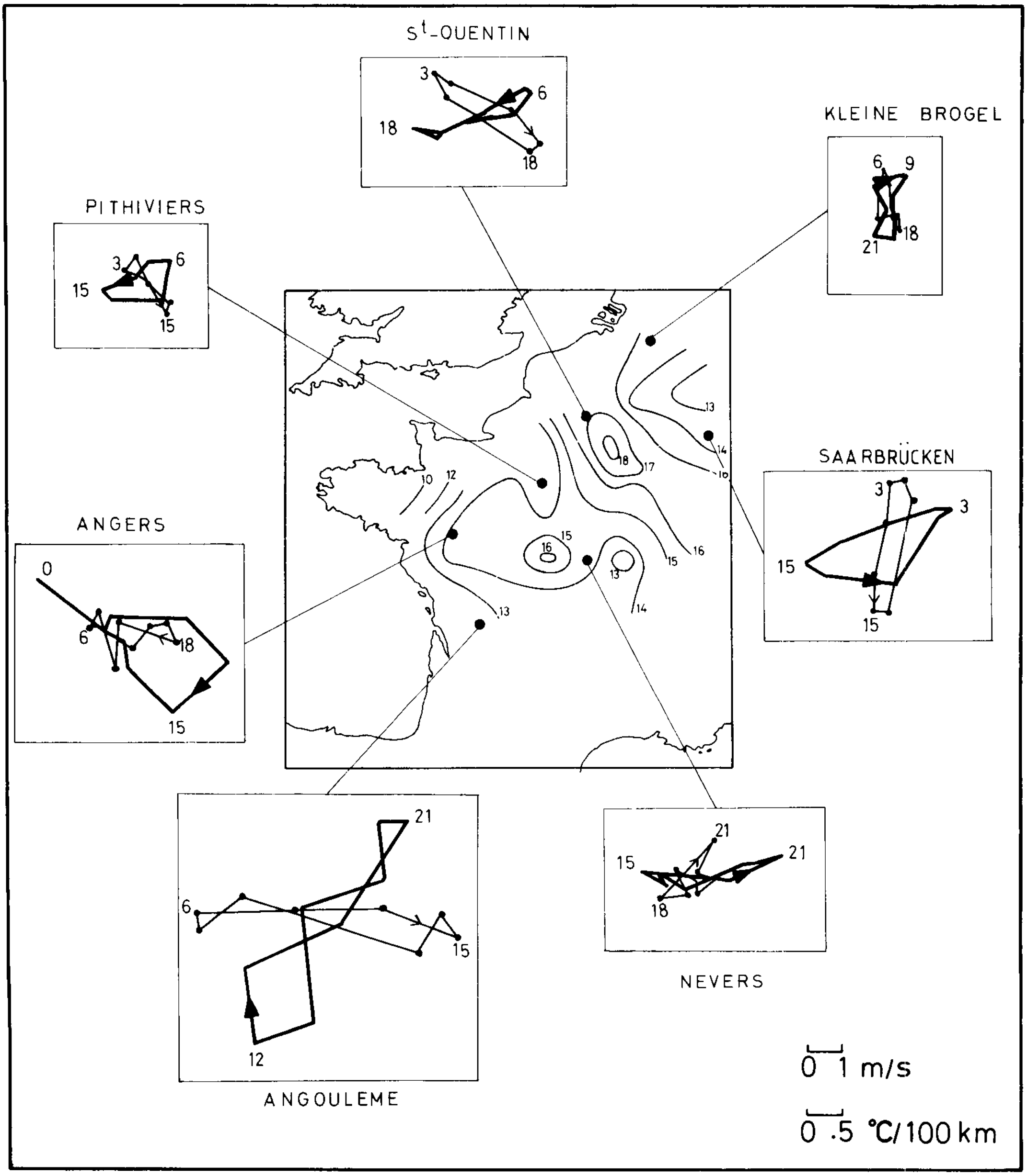

Fig. 3. Hodograph of the mean diurnal variation of the first term of eq. (2) (thick line) and of the mean diurnal variation of surface temperature gradient (thin line) (from June 28 to July 8). On each hodograph, the numbers indicate GMT. Isolines for the amplitude contours of the mean diurnal variation of the surface temperature are shown on the map. The numbers indicate the amplitude of the variation in degrees Celsius.

Paegle, 1970). The contribution of "non-topographic" effects (first term of eq. (2)) to this diurnal oscillation, however, is found to be larger in the present study than previously observed.

With the help of a rather simple theoretical analysis, some physical factors involved in the diurnal oscillation due to "non-topographic" effects have been delineated. This analysis shows that an important factor involved is the horizontally non-uniform diurnal evolution of the thermal field in the atmospheric boundary layer, which may be characterized by the geographical variations of the amplitude of the daily temperature cycle and by the characteristic height of the diabatic heating function. These results of the theoretical investigation are confirmed by the analysis of synoptic data.

In his study, Paegle (1978) implied that the diurnal oscillation in geostrophic wind due to "non-topographic" effects might be dynamically induced by external gravity waves which are excited by the air drainage along the sloping terrain. He stated that this pressure perturbation at the surface associated with gravity waves is coupled with a diurnal divergence field. In our case, the existence of the diurnal divergence field has been confirmed experimentally (Pinty, 1980), but its contribution to the surface pressure perturbation (estimated from the numerical figures used by Paegle) appears too small to be involved in the diurnal oscillation of the surface geostrophic wind. However, the relative rôle of gravity waves could be increased by considering other physical mechanisms such as a rapid standardization with height of the horizontally non-uniform thermal field, an increase of the divergence through the boundary layer, and the reduction of the effective geostrophic oscillation due to the geostrophic adjustment of the total force field.

A comprehensive evolution of the rôle of such factors requires appropriate upper air data and involves a convenient theoretical interpretation which is under investigation. Nevertheless, the present study shows that the presence of a gentle large-scale topography and the existence of a slight large-scale variation of the thermal field may significantly influence the diurnal evolution of the boundary layer, since the geostrophic wind may be considered as a dynamic forcing function in the boundary layer.

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