Microbaroms and the Temperature and Wind of the Upper Atmosphere[1]

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ABSTRACT

Microbaroms are regular pressure variations of a few microbars (dyn cm^{-2}) produced by the passage of infrasound (~5 sec period) radiated from ocean waves. Their amplitudes show prominent diurnal, semidiurnal and seasonal variations that are shown to depend on the presence or absence of one or two atmospheric sound ducts between the surface and an elevation of ~120 km. These ducts depend on the vertical temperature and wind structure of the atmosphere. For our station (Palisades, N. Y.), ducting of sound from the most common source of microbaroms (Atlantic Ocean storms) requires the presence of strong easterly winds at some upper reflection level. Variations (such as tidal) in these winds, as derived from available reports, are shown to account for the observed patterns of microbaroms. In particular, these patterns are shown to be controlled by effects of tidal and seasonal wind variations and stratospheric warmings. Having established the dependence of microbaroms on upper temperature and winds, we use the relationship to interpret these upper atmospheric conditions. Finally, we suggest that use of an expanded "synoptic" network of infrasound recorders would provide a simple procedure to monitor conditions in the upper atmosphere.

1. Introduction

Many techniques have evolved to monitor winds at heights between 20 and 120 km. By far the most reliable data come from those which measure wind motion directly, such as the use of meteorological rockets and high-level balloons for elevations below 80 km, and meteor trail observations from 90 to 120 km. The first two involve considerable expense, and the third is just now being used regularly (Revah, 1969). Indirect determinations of winds above 70 km have been made from noctilucent- and metallic-cloud drifts, the latter released by projectiles. Also, winds (the motion of neutral particles) above 60 km have been estimated from ionospheric drifts observed by the use of radio and low-frequency reflections. Many problems still exist in the interpretation of ionospheric drift observations, but with the use of certain assumptions results have been made to agree with those from other direct and indirect procedures (Fraser and Kochanski, 1970).

Studies of sporadic E (equivalent to effective wind shear) and observations of *Sq* current strengths have been used to indicate neutral wind motion in the lower ionosphere. Results differ from those of other methods (Reddy and Matsushita, 1968) possibly because of nonlinearities in tidal winds caused by gravity waves and the difficulties in interpretation of ionospheric drift data.

Another indirect method for observing wind variations in the atmosphere, which involves recording continuously generated infrasonic waves with periods of 1–10 sec, was suggested by Donn and Posmentier (1968). These oscillations, recorded as pressure variations with amplitudes of a few microbars, are known as microbaroms; their origin has been explained by generation from effects of interfering ocean waves in marine storm areas (Posmentier, 1968). This paper will explore more completely the relationship between microbaroms and upper atmospheric winds in order to show that infrasound may be useful in monitoring conditions in the upper atmosphere.

One advantage of this method arises from the ability to monitor continuously. Other methods generally provide instantaneous observations of air motions which, according to Kochanski (1964), contain a 40% contribution of gravity waves with periods of a few hours, 40% general drift, and 25% tidal components. Thus, various wave-like components recorded at a particular time may be unrepresentative of the mean air motion. Continuous recording of infrasound yields observations which effectively average out shortlived perturbations. However, as microbarom observations do not directly give information about upper atmospheric winds, interpretation is necessary. Hence, we must first review briefly the nature of microbaroms and then examine the conditions that affect their propagation;

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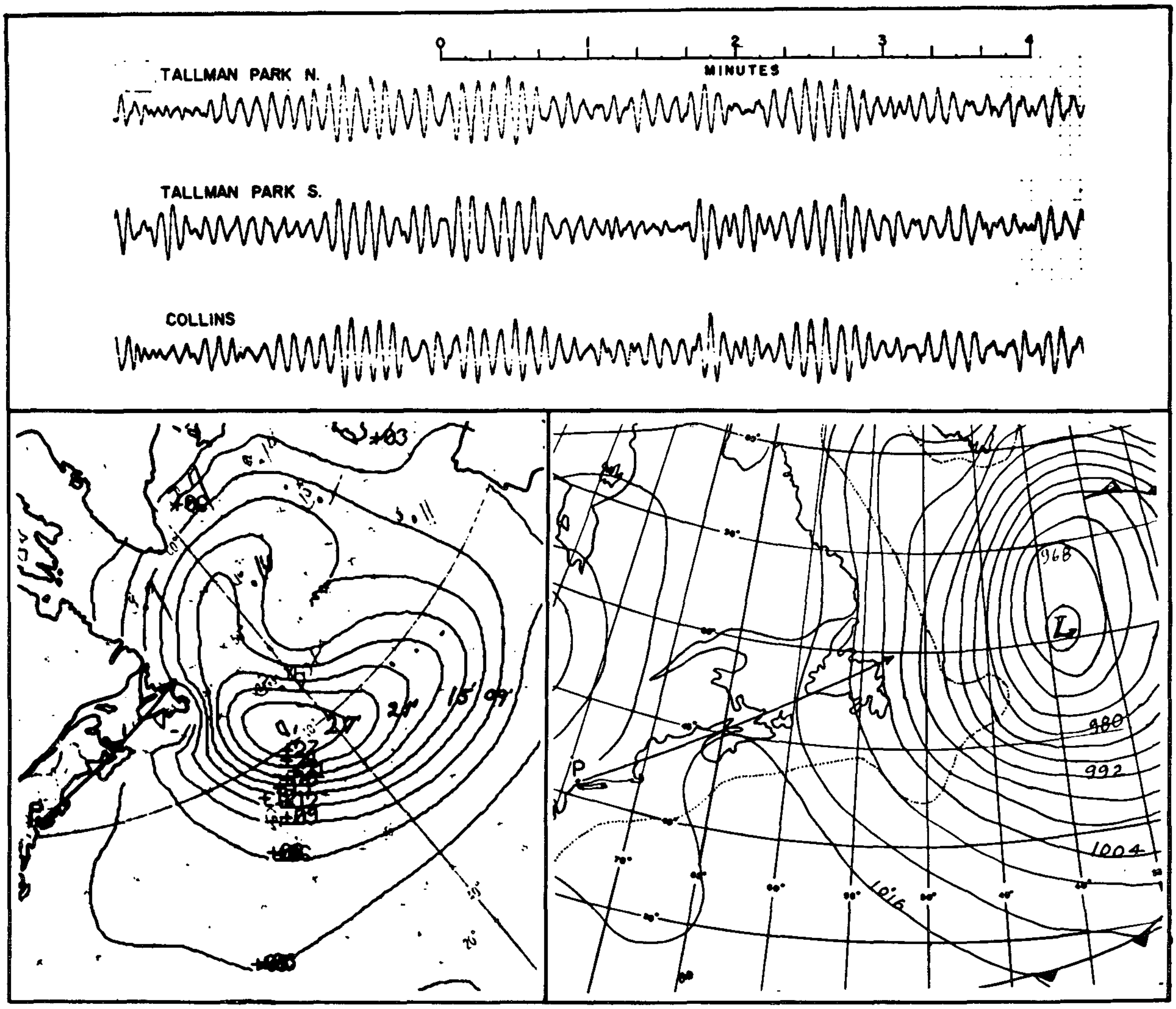


FIG. 1. Microbaroms recorded 7 April 1966 at the three elements of our tripartite microphone array (top); a coherence of 0.80 exists. Wave arrivals, which are from the source direction indicated by the long arrows in the lower charts, indicate the marine disturbance east of Newfoundland. The chart in the lower left gives wave height contours (ft) generated by the cyclone in the lower right. "P" indicates Palisades, the recording site.

these conditions can then be applied to the interpretation of microbarom observations.

2. Microbaroms

Microbaroms are infrasonic waves of about 5-sec period usually recorded as pressure perturbations of a few microbars (dyn cm^{-2}). They have been shown (e.g., Donn and Posmentier, 1967, 1968) to radiate from marine storms where they are generated as noted above by the effects of ocean waves on the atmosphere. In Fig. 1 an example is given of microbaroms recorded at Palisades, N. Y.; these have been determined to originate in the region of the marine storm shown. For the past four years we have recorded microbaroms continuously with the use of a tripartite array of capacitor microphones. Time lags among the units of the array enable the computation of horizontal trace-velocity vectors that indicate the sound source, often storms as distant as 2000 km. Details of the recording system and analytical procedure have been given by Donn and Posmentier (1967).

3. Propagation of microbaroms

The sound velocity at any level is a function of both temperature and wind, as $v=c\pm w$, where c is the sound speed dependent on temperature and w the wind speed $(=|\mathbf{w}|$, where $\mathbf{w}$ is the horizontal component of wind velocity at each elevation considered). In this study, c is expressed by a form of Laplace's equation for an ideal gas, i.e.,

$$c=(\gamma RT/M)^{\frac{1}{2}},$$

where γ is C_p/C_v, the ratio of specific heats at constant pressure to constant volume, R the gas constant, T the absolute temperature, and M the molecular weight of air. Horizontal gradients of temperature and wind are taken to be negligible for the relatively short distances involved in this study. Whereas the effect of temperature on sound velocity is isotropic, the effect of wind introduces a directional dependence.

Although horizontal gradients of temperature and wind are assumed to be insignificant on the scale of interest, significant vertical variations of these parameters exist. These vertical variations produce atmospheric sound channels or ducts. For our purposes the earth's surface may be regarded as the lower boundary of a channel; some upper level (depending on the ray angle) where the sound velocity v equals the velocity at the surface, forms the upper boundary. Hence, any signal that originates at the surface will propagate upward until reflection occurs for those rays exceeding the critical angle. Microbaroms originating over the ocean may thus be received at distant ground-based stations after one or more upper level reflections. If an adequate and continuous microbarom source is present, short-interval variations in signal strength can be related to variations in temperature or wind or both, at the reflection level.

4. Source conditions

In interpreting variations in microbarom amplitudes in terms of variations in the upper atmosphere, we must in a sense filter out possible changes in source intensity. Fortunately, an independent method of determining source conditions exists. This is made possible by the simultaneous generation of microseisms within the storm. Microseisms are seismic waves having the same period spectrum as microbaroms and have been shown to be generated by pressure variations produced on the sea bottom by the effect of standing ocean waves (Longuet-Higgins, 1950). Posmentier (1968) extended this explanation to the generation of microbaroms from the upward pressure effect of standing waves. Because microseisms propagate through the immobile crust of the earth, their speed of propagation in a particular direction is constant. Hence, variations in microseisms from a particular source reflect variations of the source itself, and provide a good control of generating conditions for microbaroms originating simultaneously from the same source. Identification of source similarity is made on the basis of a similar propagation direction for microbaroms and microseisms. Once the nature of the source is established, we propose that variations in microbaroms, not source-connected, can shed light on temperature and wind variations in the upper atmosphere. We will first review the "known" variations of temperature and wind and then interpret microbarom observations in terms of them. Once the relationship is established, the procedure can with care be reversed; variations in microbaroms may then be used to indicate temperature, and especially wind, fluctuations in a reflection zone.

5. Temperature-sound structure of the atmosphere

The speed of sound vs elevation (based on temperature alone) is given in Fig. 2 using data from the *U. S. Standard Atmosphere Supplements 1966*. On the basis of temperature, only one permanent reflecting layer exists aloft; this is above 110 km where sound speed matches that at the surface. Short-term temperature

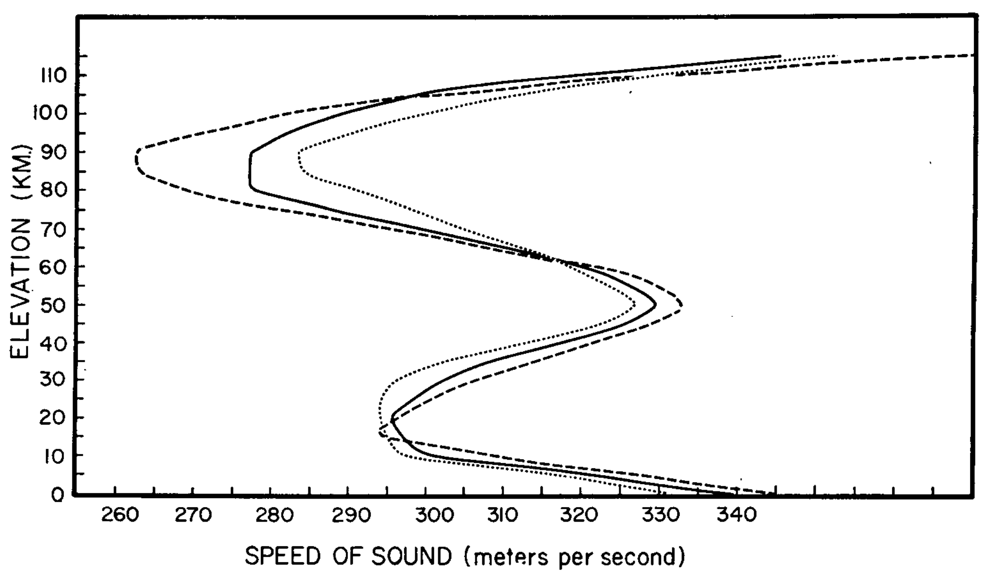


FIG. 2. Speed of sound vs altitude at 45N (from *U. S. Standard Atmosphere Supplements 1966* as a function of temperature, for January (dotted line), fall-spring (solid line), and summer (dashed line).

fluctuations and winds can modify this picture so that at times the 50-km region becomes a second reflection level.

From the tropopause to 60 km no theoretical basis for any strong diurnal temperature fluctuations has been found (Lindzen, 1967). However, observations generally show variations $\geqslant$10C at about 50–60 km (e.g., Beyers *et al.*, 1966; Smith *et al.*, 1968; Quiroz *et al.*, 1963). After all probable instrumental errors are removed, the diurnal variation is still present (Ballard, 1967; Ballard and Rofe, 1969). Observations show that in mid-latitudes, in summer especially, daytime temperatures are higher by about 10C (Beyers *et al.*; Finger and Woolf, 1967; Webb, 1966). In winter this variation is less regular and higher temperatures may occur at night (Theon *et al.*, 1967; Beyers and Miers, 1965).

No agreement exists about daily temperature variations in the region from 60 to 125 km [see, for example, Fig. 24 in Revah (1969) and Fig. 4 in Rawer (1970)]. Much of this difficulty may lie in thermal effects from viscous dissipation of gravity waves (Smith *et al.*). Also, at about 100 km there are many serious problems which complicate the measurement of temperature.

Contrary to uncertain variations of temperature, very strong seasonal and diurnal variations of wind velocity are known to occur above the tropopause. For the most part, temporal changes in microbarom signal strength will be shown here to result from variations in upper level winds rather than temperature.

6. Winds in the upper atmosphere

The prevailing wind in the region of the upper atmosphere of concern here (from the tropopause to 125 km) is strongly modulated by tidal components of varying strength and phase. Above 80 km, additional nonlinear contributions from long-period gravity waves also become important. Each of these wind patterns must be reviewed in order to make apparent their contribution to the total wind and related sound channels.

a. Prevailing wind

In Fig. 3 (after Batten, 1961), a summary is given of the monthly variation of the prevailing zonal wind in middle latitudes. The most obvious effect is the seasonal reversal of the zonal wind between 30 and 80 km. The strong west winds in winter are a consequence of the cold polar stratosphere. This mean picture can be altered significantly, particularly in mid- and high latitudes when stratospheric warmings occur between late December and February.

It will become evident later in this paper that a detailed view of wind behavior above 80 km is required for our analysis. Between 80 and 100 km, east winds are most prominent in fall and spring, while west winds predominate in summer and become stronger with increasing elevation. In winter, west winds weaken with increasing elevation (Spizzichino, 1970; Müller, 1966).

Above 100 km, existing evidence indicates that the prevailing wind is from the east in fall and winter and reaches a maximum between 105 and 110 km (Hines, 1966; and C. Justus, personal communication). Above 115 km, the prevailing wind may return to westerly (Hines, *loc. cit.*).

b. Diurnal tidal wind

According to theory (Lindzen, 1967), the diurnal tide is largely a thermally-induced oscillation caused by heating of water vapor in the troposphere from direct absorption of insolation and by ozone absorption in the stratosphere and mesosphere. In low latitudes (below 45°) heating of water vapor should lead to vertically propagating modes. In higher latitudes, ozone absorp

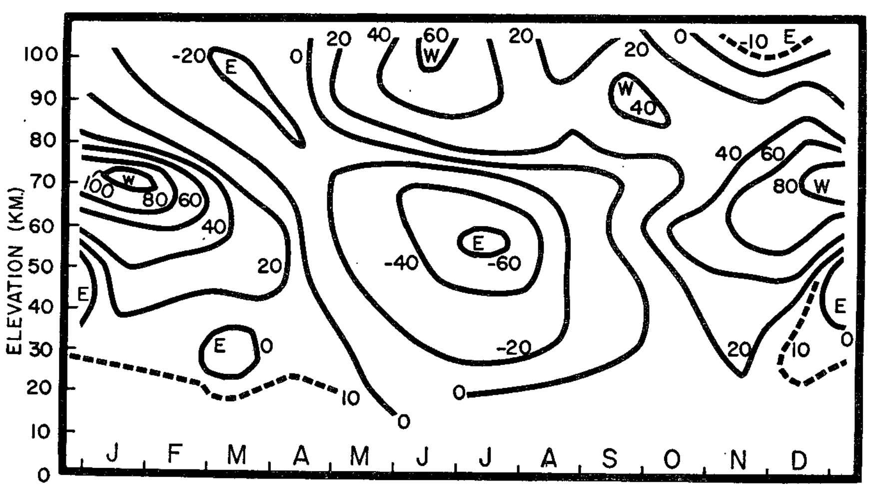


FIG. 3. Monthly variation of the prevailing zonal wind speed (m sec^{-1}) in middle latitudes (after Batten, 1961).

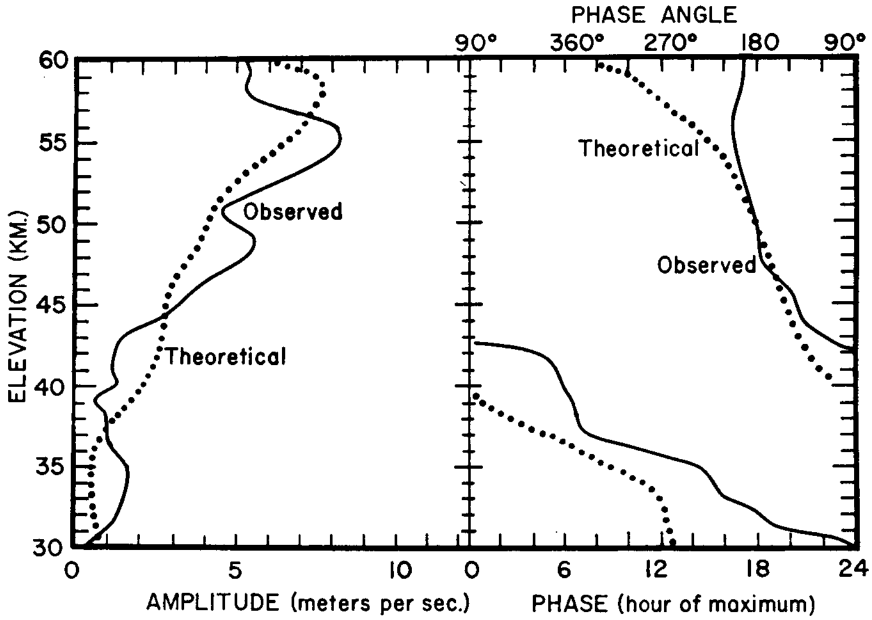


FIG. 4. Amplitude and phase of the diurnal variation of zonal wind component at 30N (after Reed *et. al.*, 1969).

tion becomes more important and leads to trapped modes that decay exponentially above 80 km. Resulting diurnal tidal winds should thus have a strong latitudinal dependence. In low latitudes, the wind should both increase and exhibit a change of phase with elevation. In high latitudes the winds should decrease above 80 km and show a constant phase.

Fig. 4, after Reed *et al.* (1969), compares theoretical with observed zonal diurnal tidal winds at 30N. Below 50 km, theoretical and observed amplitudes agree closely as does phase. Above this level, the observed phase is more stationary. Up to 60 km, the observed diurnal wind is a regular feature that shows little fluctuation.

The amplitude of the observed diurnal zonal component decreases above 80 km, especially at high latitudes (Fig. 5). Wind studies at latitudes near 45N (e.g., Spizzichino, 1969) show that the diurnal tide contributes

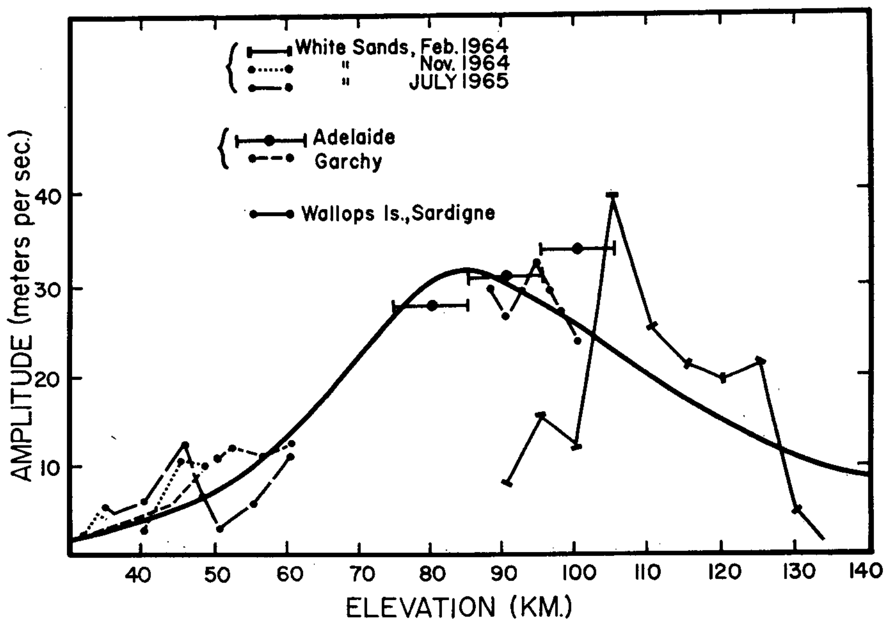


FIG. 5. Variation with height of the zonal component of the diurnal tidal wind (after Spizzichino, 1970).

little to the total wind energy above 85 km, contrary to theoretical expectations. He interprets this as a consequence of nonlinear effects of gravity waves. Thus, above 80 km the diurnal tide is weak and irregular at high latitudes; observations of Justus at 32N (personal communication) and Finger and Woolf (1967) at 35S indicate greater regularity and strength (20–40 m sec^{-1}). Our station (41N) lies between these zones.

c. *Semidiurnal tidal wind*

According to theory, this component results mainly from the heating of ozone in the stratosphere and to a lesser extent, the heating of water vapor in the troposphere (Butler and Small, 1963). Below 60 km, this component is weak and unimportant, with amplitudes <10 m sec^{-1} (Miers, 1965; Beyers *et al.*, 1966).

According to theory, the semidiurnal tidal wind ought to grow exponentially with increase in elevation above the stratosphere (e.g., Taffe, 1969). Observations by Spizzichini (1969) show that exponential growth does occur between 80 and 110 km (except in summer). In this region of exponential growth the semidiurnal wind is the most regular of tidal components, having relatively constant amplitude and phase from day to day at a given elevation. However, phase retards with a decrease in elevation between 80 and 110 km at about 7° km^{-1} in winter and 5° km^{-1} in summer. Except in summer, a 3-hr phase difference is observed between zonal and meridional components.

The observations described coincide uniquely with properties of the theoretical S_2^2 mode (in the notation of spherical harmonics where S refers to the solar-induced thermal tide) in all seasons except summer. In the latter season a weak and more variable effect results from the superposition of additional modes (Spizzichino, 1969).

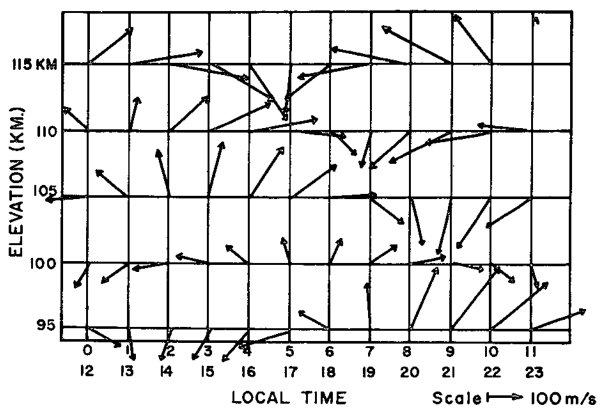


FIG. 6. Semidiurnal tidal wind as a function of hour and elevation (95–115 km) for Yuma, Ariz. (32N, 112W), 18–19 November 1967. Vectors represent direction toward which horizontal wind is moving (after Justus, personal communication).

TABLE 1. Atmospheric wind features to 125 km.

Feature	Range of importance (up to 125 km)	Comments
Prevailing wind	0–125 km	For region 30–80 km: reverses from east in summer to west in winter with speeds up to 100 m sec^{-1}. For region above 100 km: appears to be west in summer, east otherwise.
Diurnal tidal wind	0–100 km	Of major importance between 50–90 km, with magnitudes up to 40 m sec^{-1} (uncertain). Magnitudes greater at lower latitudes (below 40N).
Semidiurnal tidal wind	60–120 km	Most regular tidal component above 80 km, magnitudes up to 100 m sec^{-1}. Weaker in summer.
Gravity waves, nonlinear interactions	70–125 km	Interactions in the region of importance may be responsible for weakening of diurnal and semidiurnal tidal winds.

The semidiurnal tidal wind appears to be consistent in space as well as in time. Spizzichino compared many years of observations from six European stations and shows that a striking similarity exists in amplitude and phase among these stations. Justus (personal communication) collected a few days of data for Yuma, Ariz. These observations, reproduced in Fig. 6, show amplitudes slightly higher and phases comparable to those from Europe. Our location (near New York City) lies approximately midway between Yuma and the European stations. In a separate study, Rosenberg and Justus (1966) found that the behavior of the zonal component in phase (relative to local time) and amplitude would be consistent over a distance of at least 4000 km and the meridional component over a distance of 10,000 km.

Above 110 km, observations indicate that amplitude decreases (Spizzichino, 1970) instead of growing exponentially as predicted by theory. Spizzichino interprets this as a consequence of interaction between the semidiurnal tide and gravity waves.

A summary of the features of high-level winds referred to here is given in Table 1.

7. Observed variations of microbaroms

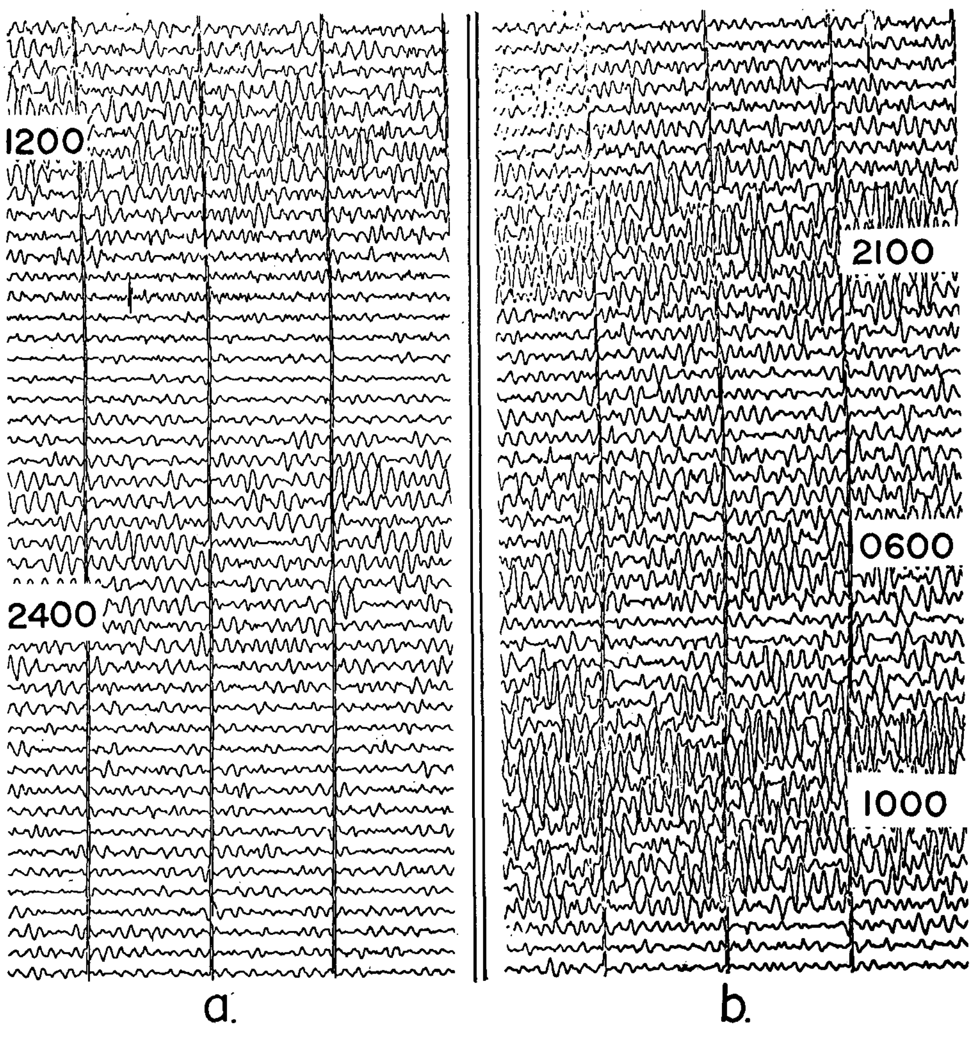
Having described the changing vertical structure of temperature and wind in the atmosphere, we can now apply this knowledge to an interpretation of observed variations of microbaroms; in the process we will note how microbaroms provide a feedback of information which permits us to determine winds. The most characteristic variations of microbaroms in the fall, winter and early spring is a semidiurnal fluctuation in amplitude as represented by the portion of a 24-hr drum record shown in Fig. 7a. At times this common pattern is altered through the occurrence of a third interval of maximum signal as in Fig. 7b. Less frequently, the semidiurnal variation is masked or nearly masked by the presence of unusually strong and diurnally continuous microbaroms as in Fig. 7c.

A fairly typical summer-type record is shown in Fig. 7d. The signal is clearly weaker than those that occur in other seasons. None of the systemmatic diurnal variations in amplitude of microbaroms are related to the behavior of the source. In addition to the fact that storms simply do not show systematic diurnal variations we know from actual observation that microseisms generated by the storms show no diurnal amplitude changes indicative of fluctuations in storm intensities. Microseismic data taken during the time of the prominent semidiurnal effect in Fig. 7a are shown in Fig. 8 which indicates essentially constant amplitude. This indicates that the storm and resulting waves had little significant change of intensity in the same interval. Such observations, involving constant-amplitude microseisms and variable microbaroms are typical of all cases studied.

8. Analysis of the usual fall, winter and early spring microbarom pattern

To investigate any systematic variations of microbaroms, measurements were made to determine their average hourly amplitude variations for several years of recordings. Results are shown in Fig. 9 for the months of October, January and April. Semidiurnal maxima are evident at about 1100 and 2300 with intervening minima at about 0700 and 1700 (all times local standard).

The semidiurnal variation is not a symmetrical one because the night maximum is clearly different from that occurring during the day. The curves show that the maximum at 1100 is more limited in duration and much smoother in appearance than the midnight peak



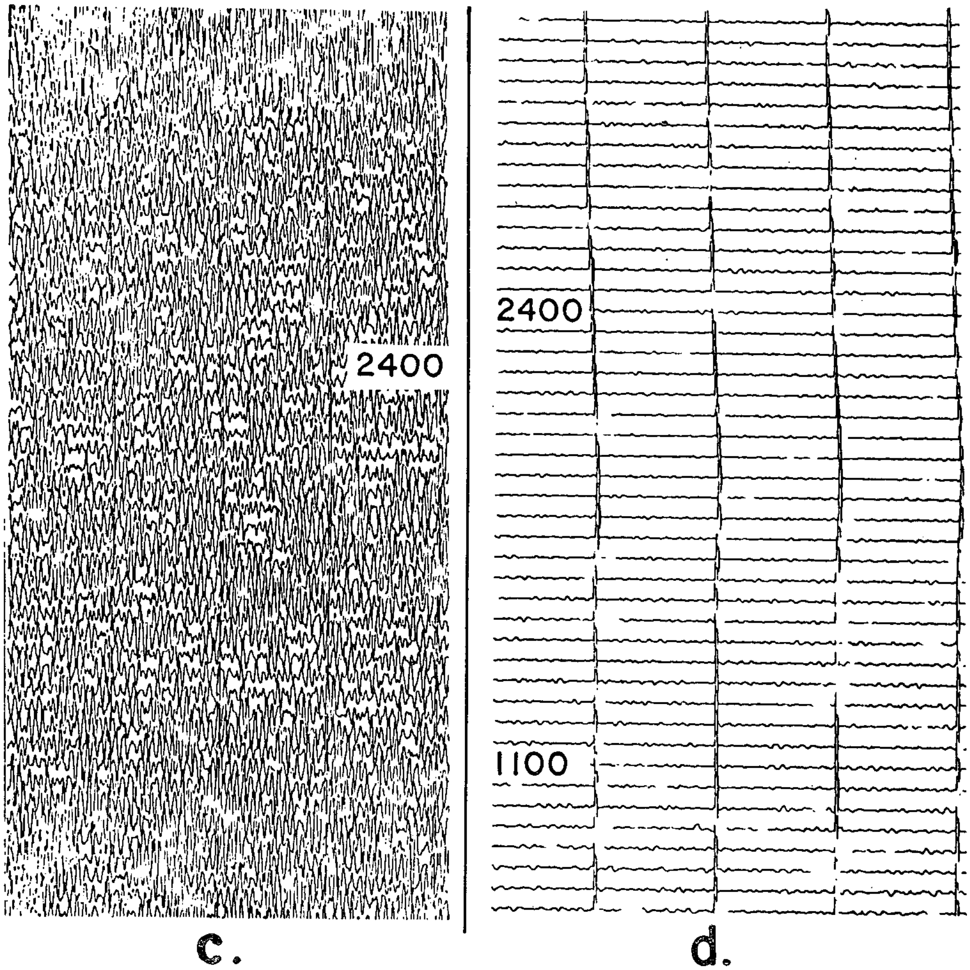


FIG. 7. Portions of microbarom drum records for 6–7 October 1969, a., 1–2 January 1968, b., 26–27 December 1967, c., and 7 September 1969, d. In each case on the original one full line represents 30 min, with a minute between time marks, with 3–4 min being shown in each half hour. In a., the semidiurnal amplitude oscillation is quite evident; in b., three separate intervals of maximum amplitude can be seen; in c., strong signal is observed continuously throughout the day; while in d., weak signal is observed.

The nocturnal maximum is considerably broader, a consequence of the averaging process which includes the less-frequent third peak at 0400, shown, for example, in Fig. 7b. In the averaging, this peak is revealed as a prolongation of the midnight effect or as a secondary, somewhat separate maximum, as in the April curve.

As the variations of microbaroms are independent of the source, they must depend on propagation factors, namely, the vertical temperature-wind structure of the atmosphere. To apply temperature-wind data to infrasound propagation, we employed a modified version of the acoustic ray-tracing program of Pierce (1966). In regular use we have found this program to be very accurate when sound source and reception points are known.

Input wind data for an average winter day as compiled from sources described above are shown in Fig. 10 for times of characteristic microbarom maxima and minima. When wind data are combined with the January temperature structure in Fig. 2, the total acoustic structure leads to the propagation patterns in Figs. 11, 12 and 13. Sound rays from a point source in the generating area were computed at vertical angle intervals of 10° between 1° and 81°.

The strong west winds around 50 km (Figs. 10a,b,c) prevent this level from forming the top of a sound channel for infrasound coming from surface sources to the east of Palisades. According to Figs. 11–13, the level of possible reflections ranges from 100 to 150 km. The particular reflection level varies with the angle of the ray and the varying velocity of the wind component from the direction of the source. This velocity variation is caused primarily by the rotating semidiurnal tidal component.

In comparing ray tracings (Figs. 12 and 13) for times of maximum and minimum microbarom amplitudes,

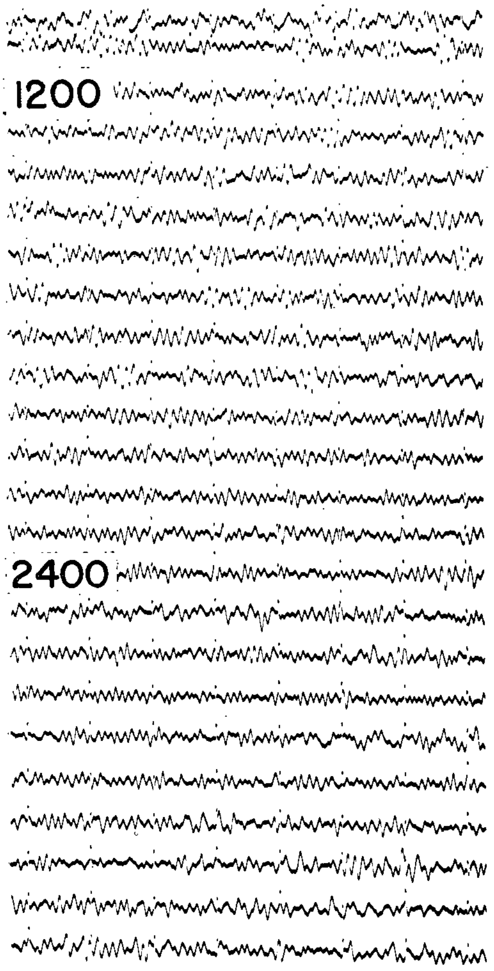


FIG. 8. Microseism record at Palisades for 6–7 October 1969. Constant amplitude is seen throughout the day. (The time scale is one-half that of the microbarom records giving only 24 lines per day.)

we note that: 1) no difference exists for rays with angles of 1°–51° whose reflection levels are above 125 km, and 2) at 1100 and 2300 rays at 61°–81° reflect between 102–108 km while the rays are reflected between 116–120 km at 1700 and 0700. The difference in amplitudes from maximum to minimum, 1100 to 1700, thus seems to be related to dissipation effects between 108 and 116 km. At 0400, a time when the intervening maximum may be present, an intermediate reflection pattern occurs. Rays at only 71°–81° reflect from 98–104 km with other reflections being from above 120 km.

To interpret the results we must consider the dissipation of acoustic energy of about 5-sec period in the elevation range of the reflections.

In a non-turbulent atmosphere, acoustic amplitudes will decrease by energy dissipation as $e^{-\alpha z}$, where the dissipation coefficient α is given by (Landau and Lifshitz, 1959; Morse and Ingard, 1968):

$$\alpha=\frac{\omega^2}{2\rho_0 c^3}[(4\eta/3)+\zeta]+\frac{\omega^2}{2\rho_0 c^3}\left(\frac{\gamma-1}{\gamma}\right)\frac{\kappa}{C_v},$$

where ω is frequency (Hz), ρ_0 density (kg m^{-3}), c the speed of sound (m sec^{-1}), η the coefficient of viscosity related to shear [kg(m sec)$^{-1}$], ζ the coefficient of bulk viscosity related to expansion [of the same order of magnitude as η and thus set equal to η (Landau and Lifshitz, 1959)], $\gamma=C_p/C_v$, and κ the thermal conductivity [kcal(m sec)$^{-1}$ (°K)$^{-1}$], and C_v, C_p the specific heats at constant volume and constant pressure [kcal gm^{-1}(°K)$^{-1}$]. Values were taken from the *U. S. Standard Atmosphere Supplements 1966*. Values of α were calculated for different frequencies and elevations with results shown in Table 2. For a frequency of 0.25 Hz (microbaroms) the acoustic signal suffers a negligible

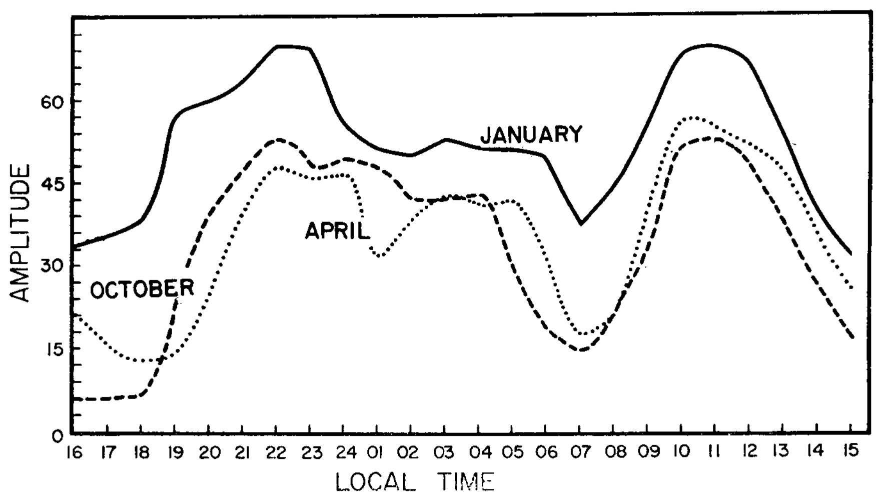


FIG. 9. Average hourly amplitudes of microbaroms for October (dashed line), January (solid line) and April (dotted line). Amplitude scale is arbitrary but consistent for the three months. Curves are drawn from three years of data for October and January, two years for April.

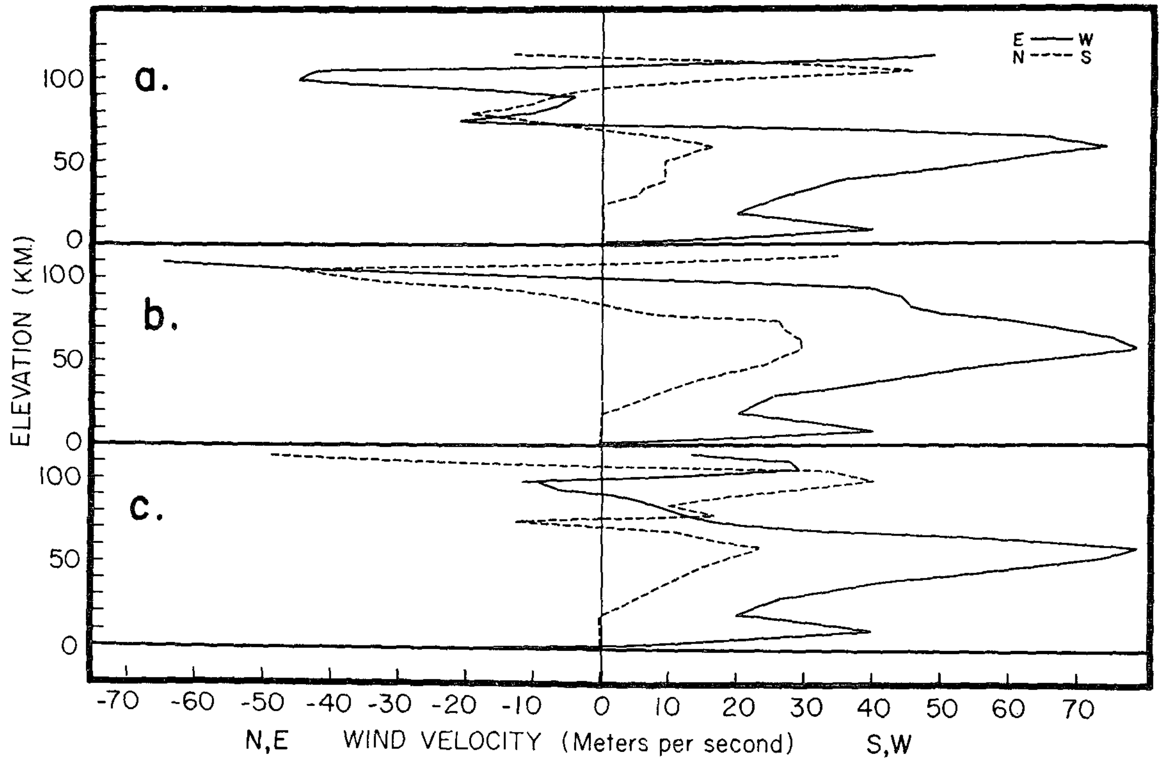


FIG. 10. Input wind data for an average winter day for 0400, a., 1100, b., and 1700, c., compiled as a "best fit" from sources given in the text.

amplitude decrease below 105 km and about 13% in traveling from 105 to 115 km. Traveling horizontally, a ray at 105 km will suffer an amplitude decrease of $1/e$ every 170 km, while a ray at 115 km has an amplitude decrease of $1/e$ every 39 km. According to our ray tracing, a ray travels horizontally about 25–30 km in the reflection region at these elevations. The resulting viscous dissipation per reflection for microbarom

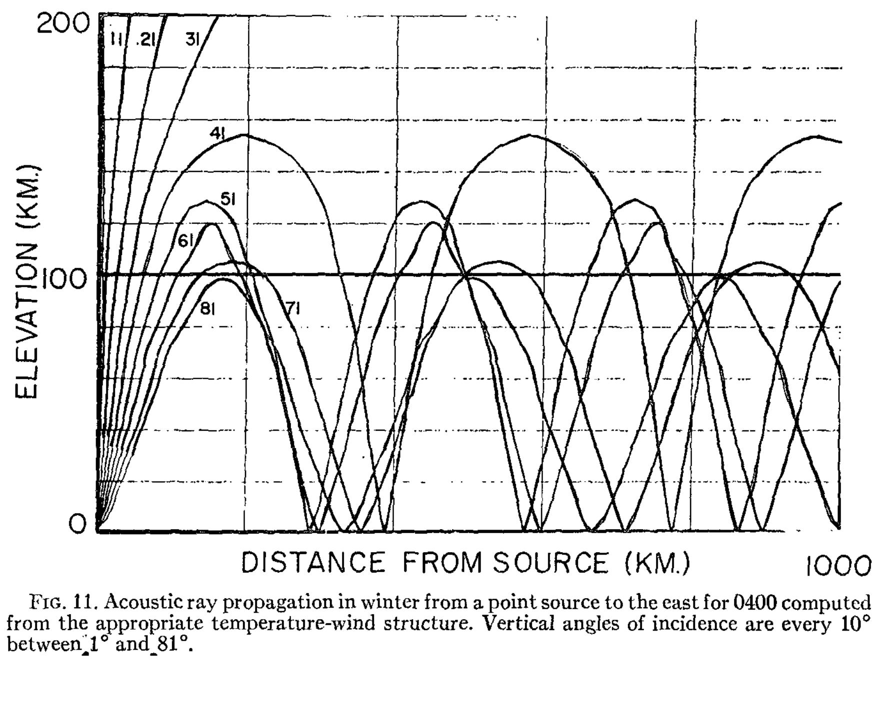


FIG. 11. Acoustic ray propagation in winter from a point source to the east for 0400 computed from the appropriate temperature-wind structure. Vertical angles of incidence are every 10° between 1° and 81°.

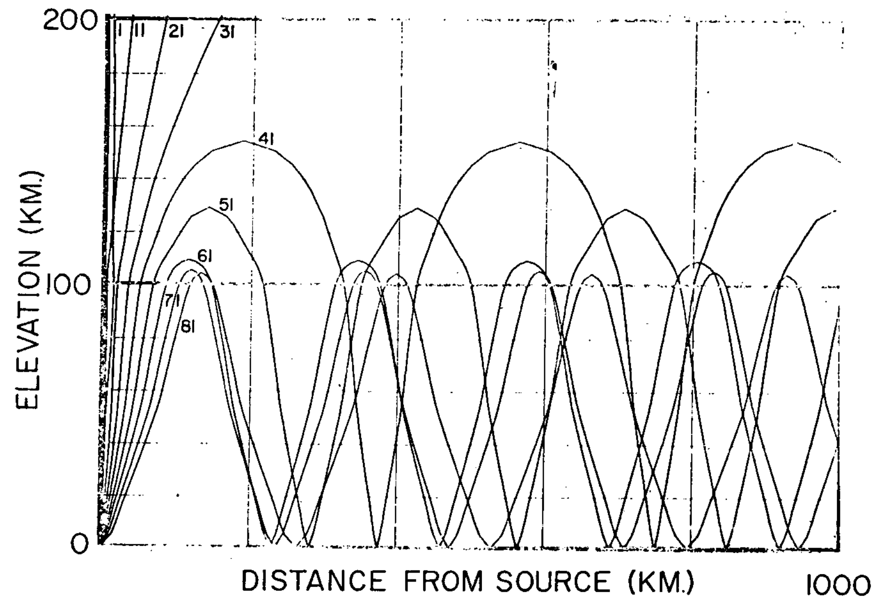


FIG. 12. Same as Fig. 11 except for 1100 (or 2200).

frequencies is about 60% greater for a ray reflected from 115 km as compared to reflection at 105 km. Because several reflections are involved, the difference in reflection level accompanied by a difference in dissipation appears to explain the diurnal variations observed. Although the equation used is strictly applicable only in the region where the amplitude decrease is small over the distance of a wavelength, and therefore of doubtful applicability around 115 km, the equation gives the lower limit of absorption at this level, and the effect would, if anything, be greater here.

There has been considerable discussion over the presence of turbulence between 90 and 120 km related to dissipation by gravity waves and wind shear (e.g., Justus and Roper, 1968; Müller, 1968; Bedinger, 1969). It would appear from microbarom reflection, concluded to occur between 100 and 115 km, that either turbulence is not a regular feature of the atmosphere at and below this region or its scale is smaller than 1 km (wavelength of infrasound); therefore, turbulent scattering of microbarom energy does not occur at levels below that where reflection occurs. Otherwise, turbulent scattering

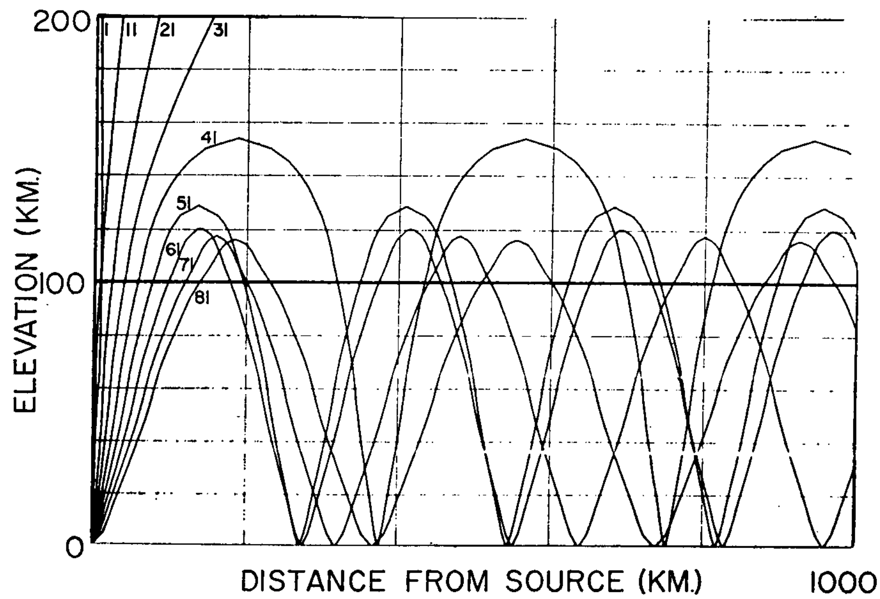


FIG. 13. Same as Fig. 11 except for 1700 (or 0700).

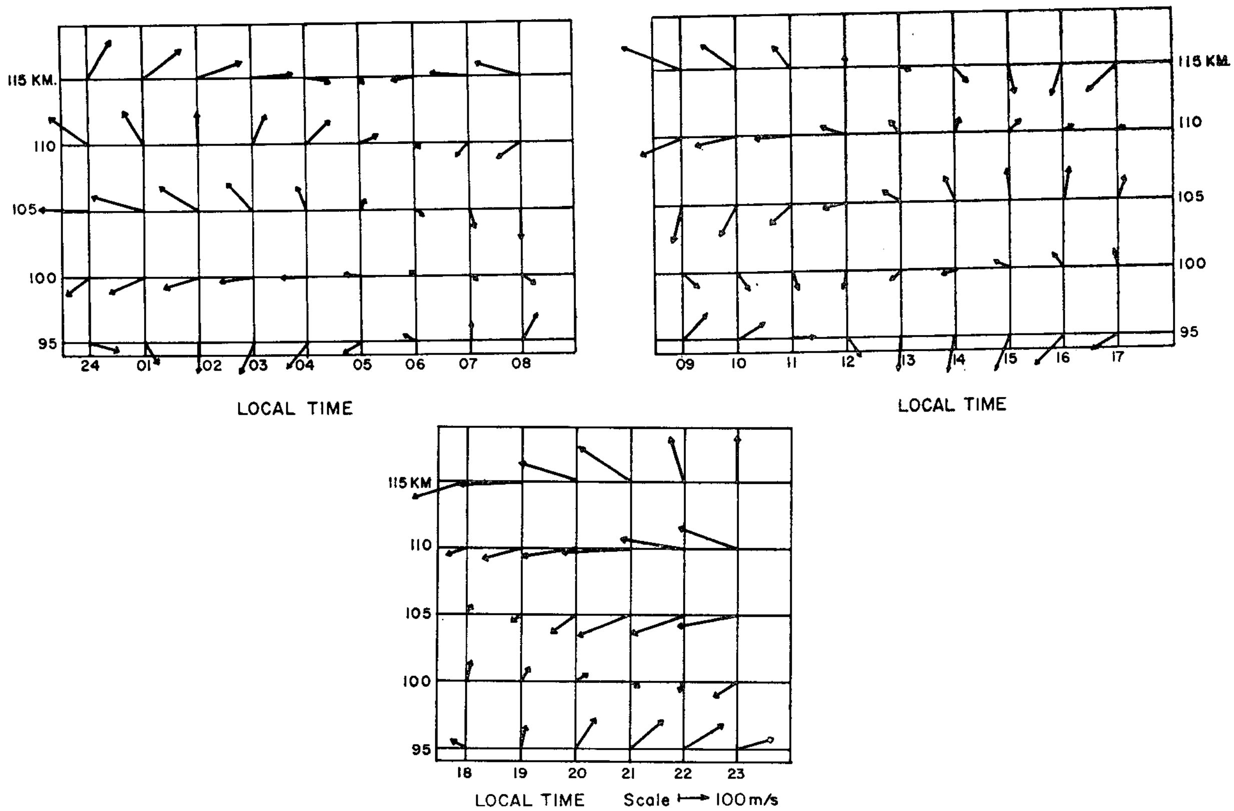


FIG. 14. Observed wind from 95–115 km for Yuma, Ariz. (32N, 112W), 18–19 November 1967. The wind vector notation is as in Fig. 6 (after C. Justus, personal communication).

of infrasound of 5-sec period would prevent propagation of the observed signal.

Although no upper wind observations are available locally, we can check our wind model compiled from theoretical and empirical sources with actual observations from several locations. The wind observed at

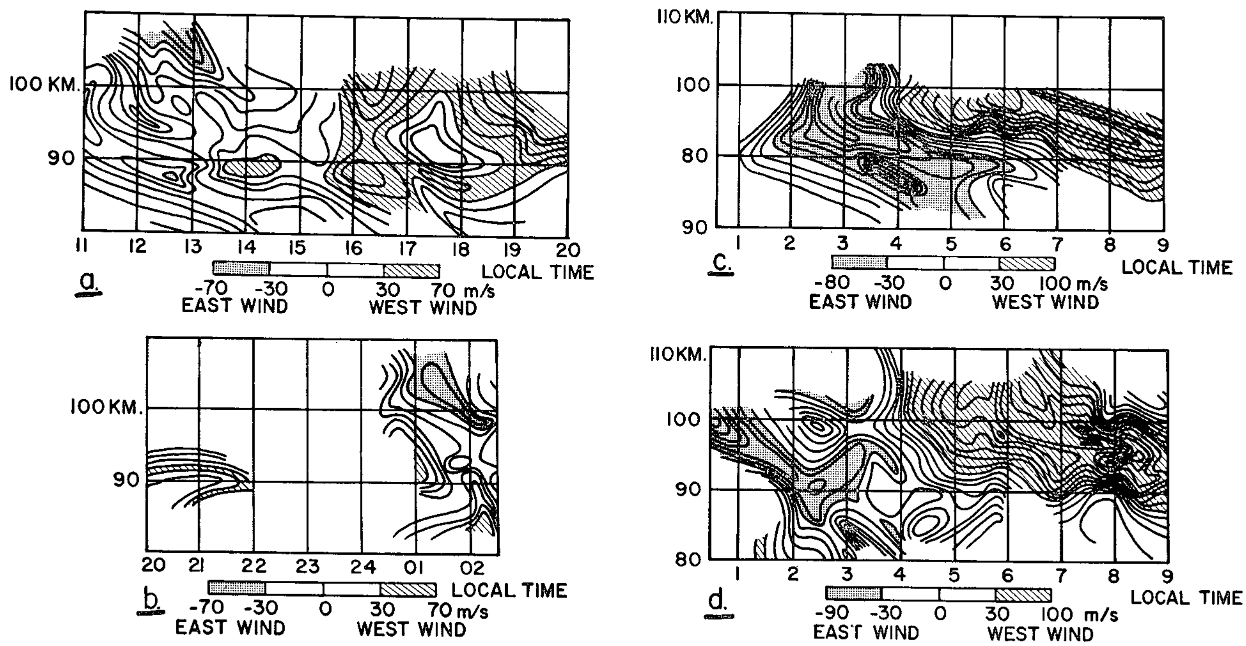


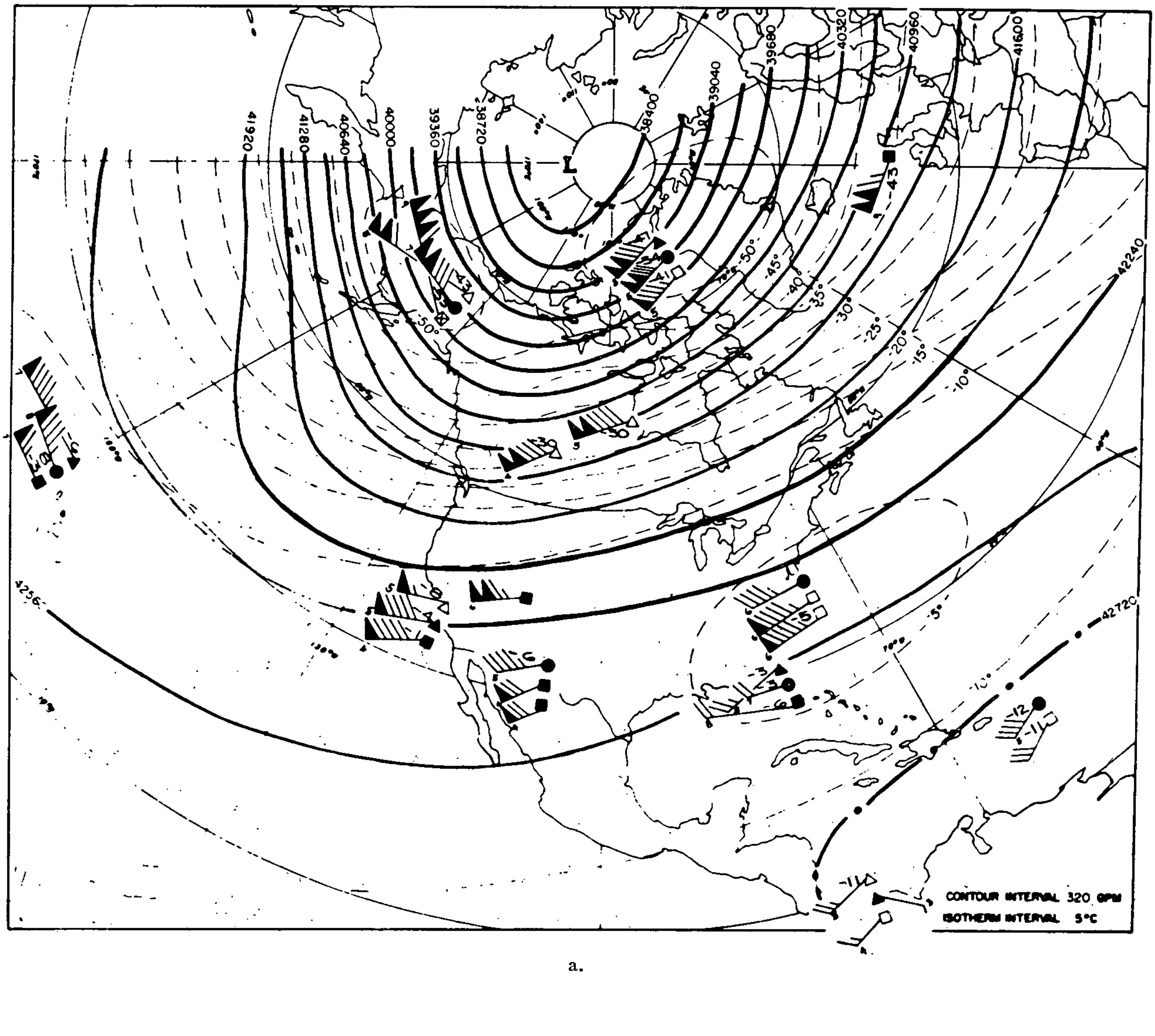
FIG. 15. Observed zonal winds (m sec^{-1}) at Garchy, France (47N), for 14 December 1965, a., 14–15 December 1965, b., 22 February 1966, c., and 23 February 1966, d. (after Revah, 1969).

TABLE 2. Dissipation coefficient (α) per kilometer vs elevation for several frequencies.

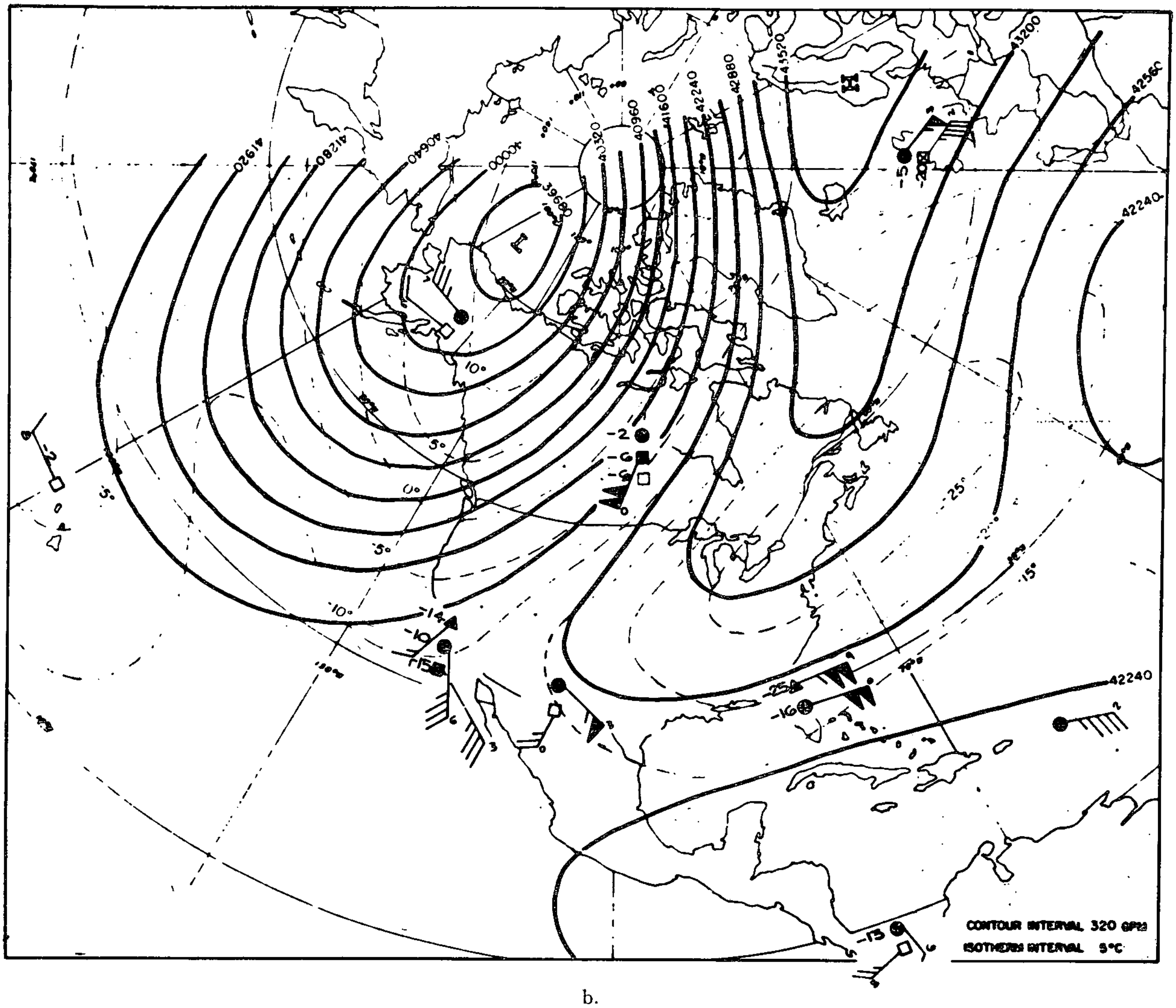
Elevation (km)	Frequency (Hz) 1	0.4	0.25	0.2
0	2.0×10^{-8}	3.2×10^{-9}	1.2×10^{-9}	7.9×10^{-10}
20	3.7×10^{-7}	5.9×10^{-8}	2.3×10^{-8}	1.5×10^{-8}
40	8.2×10^{-6}	1.3×10^{-6}	5.1×10^{-7}	3.3×10^{-7}
60	1.1×10^{-4}	1.7×10^{-5}	6.8×10^{-6}	4.3×10^{-6}
70	4.3×10^{-4}	6.8×10^{-5}	2.7×10^{-5}	1.7×10^{-5}
75	9.0×10^{-4}	1.4×10^{-5}	5.6×10^{-5}	3.6×10^{-5}
80	1.9×10^{-3}	3.0×10^{-4}	1.2×10^{-4}	7.5×10^{-5}
85	4.2×10^{-3}	6.7×10^{-4}	2.6×10^{-4}	1.7×10^{-4}
90	9.7×10^{-3}	1.5×10^{-3}	6.0×10^{-4}	3.8×10^{-4}
95	2.2×10^{-2}	3.5×10^{-3}	1.4×10^{-3}	8.7×10^{-4}
100	4.9×10^{-2}	7.8×10^{-3}	3.0×10^{-3}	1.9×10^{-3}
105	9.31×10^{-2}	1.48×10^{-2}	5.80×10^{-3}	3.71×10^{-3}
110	1.99×10^{-1}	3.18×10^{-2}	1.24×10^{-2}	7.9×10^{-3}
115	4.10×10^{-1}	6.56×10^{-2}	2.55×10^{-2}	1.63×10^{-2}

Yuma, Ariz., is shown in Fig. 14. Strong east winds are found between 100 and 110 km around 1100 and 2300, in agreement with interpretations from microbarom observations. Easterly winds of slightly weaker amplitude are reported between 95 and 100 km about 0400, the time of occasional maximum microbaroms. Although easterlies at this level are also reported at 1700, no corresponding microbarom maximum is observed, indicating the possible absence of these winds in our region.

Fig. 15 depicts the observed wind during several winter days in France at 47N. East winds can be seen between 100 and 110 km about noon and midnight; west winds prevail from 0500–0900 and 1400–2100, in agreement with the wind model used. The variability at 0400 can be seen by comparing observations for 22 February, when strong east winds are evident between 95 and 100 km from 0300–0400, with those of 23 February, when the effect was much weaker. The variability of the 0400 microbarom maximum may thus be related to the variability of winds at this time at the reflection region of about 95–105 km.



a.



b.

FIG. 16. Height contour chart at 2 mb for 6 December 1967, a. (the polar vortex with usual winter westerlies dominates mid latitudes); and 27 December 1967, b. In the latter case the polar vortex has been displaced westward as warm ridge intrudes from the Atlantic Ocean. Over the eastern United States, stratospheric easterlies now appear.

9. Analysis of microbaroms related to stratospheric warmings

The curve for January microbarom averages shows weaker minima than those for the other months. This results from several intervals of unusually high and continuous signal throughout the day as illustrated in Fig. 7c. One such interval, which lasted from 22 December 1967 to 21 January 1968, was studied in detail. Throughout most of this month microbarom activity resembled that shown in Fig. 7c unless obscured by local winds. Superimposed on the continuous signal are the microbarom variations described in Section 8.

During the above interval the circulation of the upper atmosphere was disrupted by a stratospheric warming which is evident in a comparison of the stratospheric charts (Figs. 16a and b). This warming, which has been described in detail by Johnson (1969), resulted from the movement of a stratospheric warm area in the Atlantic Ocean toward the north pole and thus affected the region critical to our microbarom observations. The warming caused a temporary reversal of the meridional temperature gradient, with the consequent destruction of the characteristic westerly circulation of the winter stratosphere.

The abnormally strong and continuous microbaroms can be explained by this changed atmospheric structure. With west winds no longer predominating in the middle stratosphere, the probable easterly wind and increased temperature structure around 50 km could establish a stratospheric sound reflector which would be below the level of tidal wind variations. Furthermore, the warming of this region over the Atlantic Ocean undoubtedly affects the semidiurnal tidal wind, a thermally induced feature whose changes could contribute further to the observed microbarom effect.

10. Analysis of microbaroms in late spring and summer

Average diurnal patterns for spring and summer are shown in Fig. 17. Note that two types of diurnal variations are indicated for the late spring (May). One (May 1968) resembles the standard winter pattern with maxima about 2300 and 1000–1100. The other is very similar to the summer pattern which features a rather prominent minimum about 2000, the time of a winter maximum.

As noted in Section 9, the winter preceding May 1968 was characterized by the presence of a strong and prolonged stratospheric warming originating over the Atlantic Ocean. Kreister (1968) found a correlation between Atlantic type warmings and the later appearance of the summer east wind circulation in the stratosphere. Hence, the difference between the May 1968 pattern and the May 1969 and typical summer pattern can be explained by the prolongation of the winter circulation into late spring during 1968.

Summer microbaroms, as shown by the July curve, are characterized by low-amplitude variations showing minimum amplitudes at about 2000. The low amplitudes in summer (as shown, for example, in Fig. 7d) are primarily related to the absence of a disturbed ocean. This is confirmed by our observations that microseisms are generally very weak in summer (as was true for the case illustrated in Fig. 7d). The significant effect in summer seems to be the presence of a diurnal rather than a semidiurnal variation. To explain this effect we again turn to the temperature-wind structure.

An important change from winter to summer is the development of steady easterly winds in the stratosphere, evident in Fig. 3 between 20 and 80 km. These winds, especially in the region of the temperature maximum around 50 km, would cause reflection of microbaroms from an easterly source. Such stratospheric reflection is not possible in winter. A striking difference between summer and winter propagation is evident in the comparison of Fig. 18 with Figs. 11–13. Reflection from the 50-km region in Fig. 18 provides the subdued signal received from the weak summer sources. The higher reflections from above 120 km, a result of the thermosphere temperature increase, would be dissipated as described earlier.

To explain the 2000 minimum, ray tracings were computed for various combinations of theoretical models with observed data. The model giving the best solution for our observations included a 12K amplitude temperature variation, reaching a maximum at noon at 50 km, and a diurnal tidal wind variation of stationary phase (along the vertical) with a maximum westerly component at 2000. For this model, ray tracings indicate a loss of sound from a vertical sector of 9° at 2000 compared to other times of the day. Although the reality of the model used is not firmly established, our conclusions support it.

11. West Coast observations of microbaroms

Because the marine microbarom source for western stations lies in the Pacific Ocean, very different effects can be expected. In winter, stratospheric westerlies should produce a sound structure analogous to summer conditions for eastern stations because in both situations stratospheric winds travel from the source region to the stations. In winter, the sound paths in Fig. 18 would apply generally to western stations.

To test these deductions we obtained observations of microbaroms in winter from Pullman, Wash., 50N, 135W (Craine and Rezvani, 1970). No obvious diurnal variation pattern is evident in these observations, indicating stratospheric reflection of most of the energy

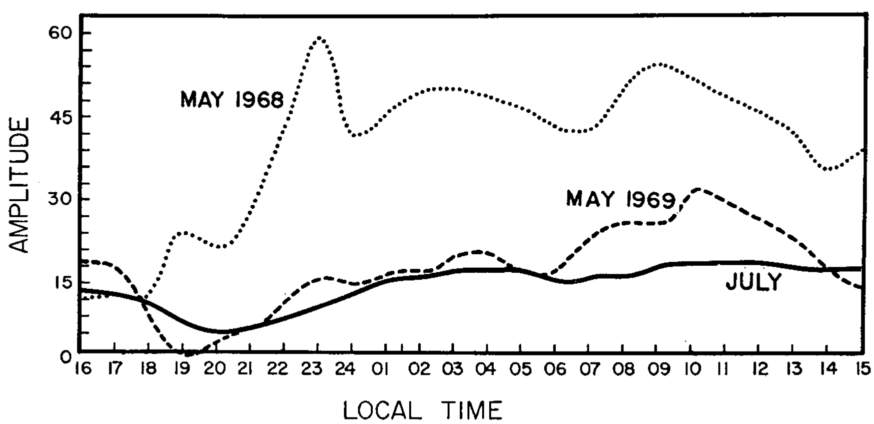


FIG. 17. Average hourly amplitude of microbaroms for May 1968 (dotted line), May 1969 (dashed line) and July from two years of data (solid line). The arbitrary amplitude scale is the same as that of Fig. 9.

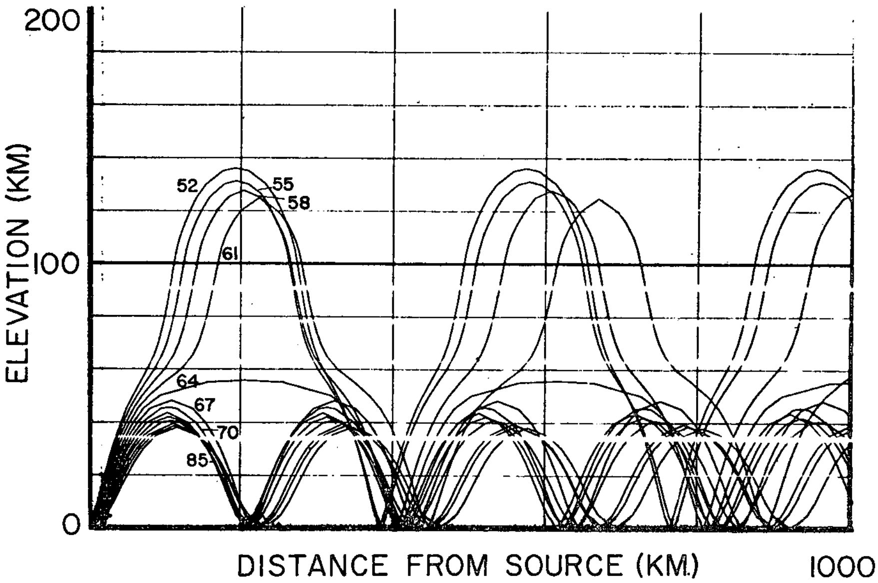


FIG. 18. Acoustic ray propagation from a print source to the east in summer at 1200 local time. Vertical angles of incidence are shown for every 3° between 52° and 85°.

by the non-rotating wind. The same effect also exists for microbaroms recorded at San Diego (Craine[4]).

12. Discussion and conclusions

Although our microbarom recording has been in continuous operation for only four years, the use of infrasound has enabled us to verify many observations and conclusions on wind variations in the region between 20 and 120 km. The apparently complicated diurnal and seasonal patterns of microbarom variations can be readily simplified and explained through the use of particular wind models. Once established, we can use microbaroms to probe and monitor much of the behavior of stratospheric and thermospheric winds. The present study has led us to make the following conclusions:

1) The semidiurnal tidal winds contribute greatly to the total winds from 90–120 km in much the way described in detail for Arizona and Europe.

2) Easterly winds near 100 km appear about 0400 in an aperiodic fashion, giving the nocturnal wind pattern a different appearance from that of the day.

3) The onset and duration of a stratospheric warming in our region is indicated by the unique appearance of strong and diurnally continuous microbaroms.

4) The beginning of the summer stratospheric circulation (easterlies) in our region is marked by the onset of a later spring-summer microbarom pattern characterized by a single minimum occurring prior to local midnight.

5) The summer nocturnal minimum seems to depend on an atmospheric model which, based on available information, includes a diurnal wind with stationary phase (along the vertical) and a diurnal temperature variation of 12K amplitude at 50 km, with maximum at noon.

Because our detailed observations refer to one region, we propose that synoptic data from a number of appropriately distributed infrasound recorders would help test these conclusions. When verified, such a system would provide a probe of the upper atmosphere through the use of readily available and continuous natural atmospheric background phenomena.

We might predict a strong latitudinal dependence of microbaroms on upper winds. In lower latitudes, where the diurnal tide is more important than the semidiurnal (Rao and Rao, 1965), a corresponding effect in the microbarom variation pattern should be apparent. And in high latitudes where the diurnal tide is insignificant, and the semidiurnal tide stronger, the semidiurnal microbarom pattern would be even more prominent than that which we observed.

In addition to the several, rather gross microbarom-wind relationships described, we have observed more subtle effects which are in the process of investigation.

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[4] Paper presented at 1970 annual meeting, American Geophysical Union, Washington, D. C.

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