Further Use of Natural Infrasound as a Continuous Monitor of the Upper Atmosphere [1,2]

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ABSTRACT

This is a further study of the use of natural infrasound in the atmosphere to monitor tidal circulation in the lower thermosphere. The height of this circulation is determined with the use of a reference atmospheric model, which is then used to calibrate infrasound/microseism ratios in terms of height. Also, we show from continuous observation over six years at 41°N, 74°W that the winter semidiurnal tide is present at least 62% of the time and the diurnal, at least 42% of the time.

1. Introduction

This paper reports the attempts to refine the infrasound technique for monitoring upper atmosphere winds and temperatures (Donn and Rind, 1972; Rind *et al.*, 1973). Naturally occurring infrasound is generated continuously by ocean waves and propagates up to the stratosphere or lower thermosphere where it undergoes reflection from the combined effects of wind and temperature. In this paper we will discuss several methods for estimating the level of reflection. We then apply these results to determine the regularity of upper air circulations, which is possible because the system is in operation 24 h a day, every day of the year.

Infrasound has been recorded for the past eight years at Palisades, N. Y. (41°N, 74°W), with the use of capacitor microphones arranged in a tripartite array. Relevant details are available in Donn and Rind (1972). Cross correlation of the data recorded both on magnetic tape and on visible chart records allows the direction and trace velocity of the signal to be determined. We have already shown (Rind *et al.*, 1973) that the trace velocity is equal to the speed of sound which is a function of wind and temperature at the reflection level. In all of this work, a crucial problem is the determination of the level of reflection; this is discussed in detail below.

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[2] University Institute of Oceanography, City University of New York, Contribution No. 58.

2. Absolute height determination

Reflection of infrasound originating at the surface occurs when the sound velocity at any level exceeds the sound velocity at the surface. The sound velocity in any direction $V(\mathbf{i})$ is equal to the component of the wind velocity W in that direction plus the sound speed C. Thus,

$$V(\mathbf{i}) = W\mathbf{i} + C, \tag{1}$$

where $C=(\gamma RT/m)^{\frac{1}{2}}$, T is the absolute temperature, R the universal gas constant, $\gamma = c_p/c_v$, and m is the mean molecular weight of air. To determine the probable level of reflection we must know the distribution of these parameters in the vertical. Thus, we are initially dependent on observations collected by other methods. For the purpose of composing such a model we conducted an extensive survey of published wind and temperature observations; the relevant sources, techniques and locations are given in Table A1 in the Appendix. The model was composed for our latitude and longitude for each month of the year; it includes prevailing effects and semidiurnal and diurnal tides for both wind and temperature at 5 km height intervals between 0 and 120 km. We used data for other latitudes bearing in mind the expected latitudinal variation of the particular phenomenon (for instance the $S_{2,4}$ semidiurnal tide). The ionospheric drift observations were used with due consideration for all the ambiguities involved in their interpretations. Finally, we compared our model with the prevailing wind models of Groves (1971) and Theon (1972) [and others listed in Table A] and

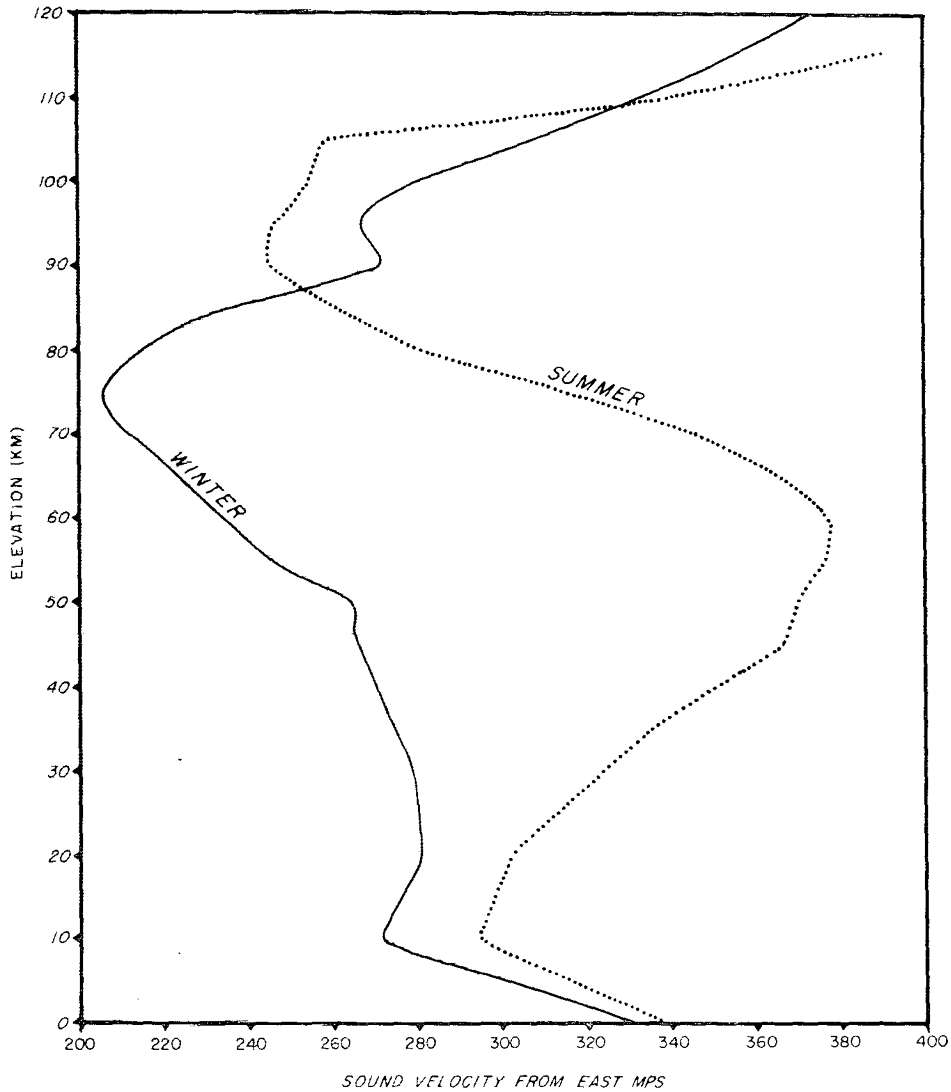

Fig. 1. Sound velocity vs height for February (solid line) and July (dotted line) computed from wind and temperature model.

mean temperature models of Champion (1967), Cole and Kantor (1972) and Salah *et al.* (1974), among others, to attempt to clarify any discrepancies.

Fig. 1 shows the model results for sound velocity from the east (where our strongest sources are located) for prevailing conditions calculated with the use of (1) for February and July (February rather than January is used to reduce the influence of stratospheric warmings in this discussion).

In summer, because the average sound velocity is greater than that at ground level from about 38 to 65 km, it is within this region that the signal will be reflected. In winter, because of the strong stratospheric west winds below, reflection first becomes possible only above 100 km. As the temperature rises strongly in the lower thermosphere, reflection should

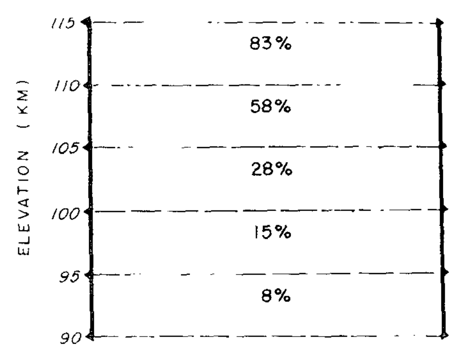

Fig. 2. Dissipation vs height for signal of 4 s period for two reflections calculated from standard dissipation coefficients (*U. S. Standard Atmosphere Supplements 1966*).

TABLE 1. Model of semidiurnal tide.

	Elevation (km) 85	90	95	100	105	110	115	120
Time of maximum west wind	1100	0930	0800	0630	0430	0300	0200	0100
Time of maximum east wind	1700	1530	1400	1230	1030	0900	0800	0700
	0500	0330	0200	0030	2230	2100	2000	1900
Velocity (m s^{-1})	15	23	31	38	47	49	49	47

be certain from above about 110 km. However, this is a region of strong dissipation for signal of the frequency recorded (Donn and Rind, 1972). Fig. 2 shows the cumulative dissipation for sound of this frequency along its path through these levels. Clearly, signal recorded from reflections above 110 km should be weak; furthermore, changes in the reflection level of 5 km should be readily apparent as changes of amplitude are dependent upon the height-related dissipation that occurs.

The exact level of reflection is, of course, determined by the actual temperature and wind at any time. This depends upon tidal components also, not simply prevailing conditions. Table 1 shows the model we synthesized for the semidiurnal tide for the region of possible reflection in winter. For comparison, one can refer to Goodwin (1968), Groves (1971), Bernard (1974a, b) and Fellows *et al.* (1974), among others. The semidiurnal tide appears to be a stable feature and there is not much disagreement about its phase at these heights. If one adds the model in Table 1 to that of the prevailing wind and temperature in Fig. 1 the following can be expected: as the tidal wind becomes east at levels close to 100 km, the signal will be reflected from these heights rather than from higher, more dissipative regions, where the high temperatures always provide a sound reflection level. The amplitude of the signal should thus show a semidiurnal variation in agreement with the tidal east wind descent evident in Table 1.

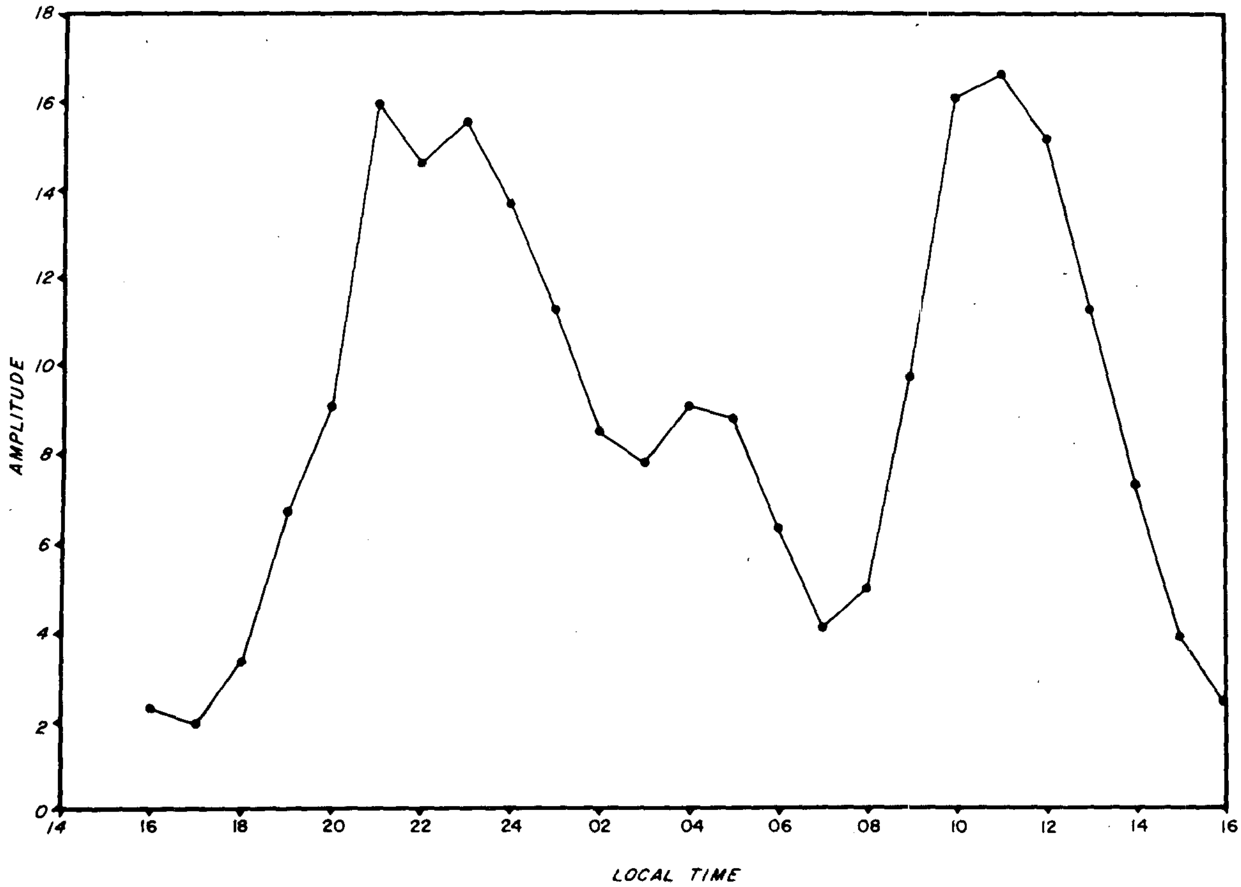

FIG. 3. Average infrasound amplitude (arbitrary units) vs time of day for February from 6 years of data.

TABLE 2. Directional rotation of infrasound.

	Hour											
	06	08	10	12	14	16	18	20	22	24	02	04
Average azimuth of signal from east	85°	92°	109°	120°	140°	115°	110°	86°	93°	107°	120°	133°
Percent signal from west	25	06	06	12	15	43	39	20	12	31	17	26

Fig. 3 shows the average February infrasound amplitude as a function of the hour of the day (in arbitrary units) from 6 years of data. It can be seen that the variation has a strong semidiurnal character with maximum amplitude at 1000–1200 and 2100–2400 local time. Comparison with the phase model in Table 1 indicates strongly that the high-amplitude signal from the east comes from heights of about 105 km.

If this effect is really due to the rotating tidal influence, it should be apparent in several ways. As the direction of the semidiurnal tidal wind rotates, the direction of the signal capable of being reflected should change very slightly for a sole narrow source but more distinctly for a broad or multi-source situation. If we average over prolonged periods, a wide scatter of sources will exist, and a directional rotation of infrasound should become evident. In Table 2 constructed from computed azimuths by averaging 345 h of data scattered over 6 years such an effect is shown. The most southerly (130°–140°) azimuth agrees with a semidiurnal meridional wind in quadrature with that shown in Table 1 for 105 km.

One final presentation of this rotational effect can be made. In Fig. 4 the amplitude variation versus time is displayed for signals from the northeast, east and southeast. The phase of the reflecting (tidal) wind responsible for such variations agrees with results shown above and fixes the average region of reflection at 105 km for the high-amplitude signal, and above this level for signal of lower amplitude. These two studies verify further that our interpretation of reflection provided by the rotating semi-diurnal tide is valid.

3. "Independent" height determination

Having determined the approximate height of reflection from data determined by other techniques we are now in a position to apply this result to develop an independent method. This method involves relating the signal amplitude to the source strength as indicated by microseisms generated simultaneously. Microseisms are short-period seismic waves radiated through the crust of the earth from the effect of inter-

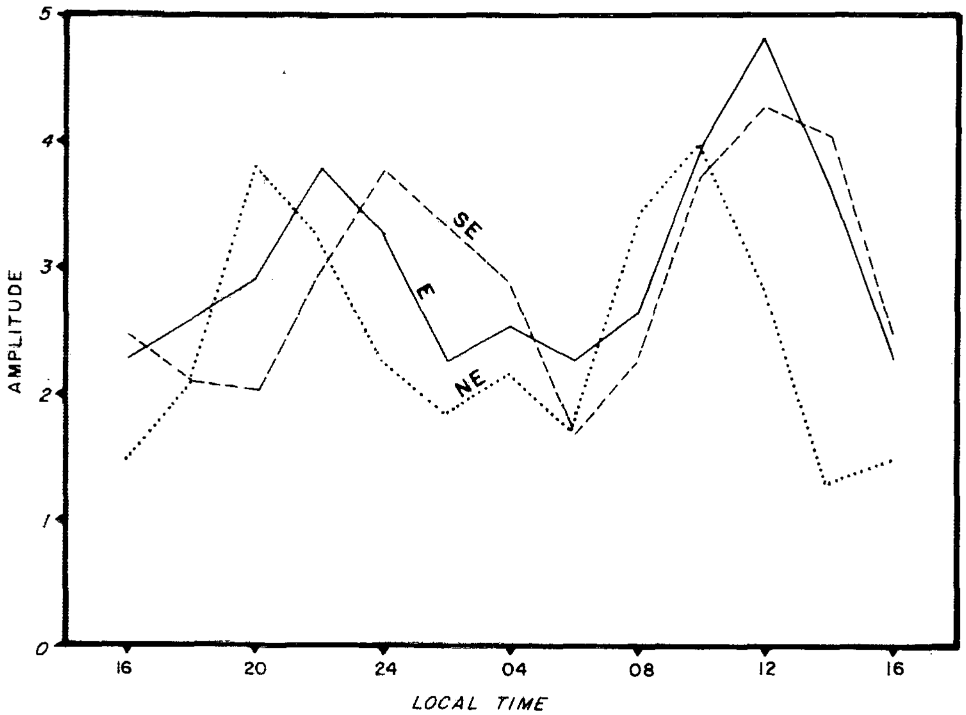

FIG. 4. Average amplitude (arbitrary units) vs time of day for infrasound from the northeast (dotted line), east (solid line) and southeast (dashed line).

TABLE 3. Infrasound parameters for various atmospheric situations.

	Winter (120 h of data) Maximum (1100, 2300)	Intermediate (0900, 2000)	Minimum (0700, 1700)	Summer (40 h of data)
Ratio (I/M)	0.4	0.2	0.1	0.54
Estimated reflection height (km)	105	110	110+	45–50
Average sound velocity (m s^{-1})	350	348	335	365
Average direction (deg)	96	90	97	105

fering ocean waves on the sea bottom. According to Posmentier (1967) and Donn and Naini (1973), the storm waves simultaneously generate atmospheric infrasound of the same period as the microseisms.

The main parameter in the "independent" method of height determination is the ratio of infrasound to microseism amplitude (I/M). We first determine this ratio for signals known, from the model described, to be reflected at 105 km with about a 5 km height resolution. As this ratio rises and falls we estimate the reflection level lowering and rising respectively below or above 105 km. This study is still in its infancy; the data presented here are a preliminary indication of what may be accomplished.

At Lamont, standard recording of the three components of ground motion (vertical, east-west and north-south) is made. Because microseisms contain a dominant Rayleigh wave component one can correlate the different ground motions to determine the direction of the signal. This has been done using an analogue correlator. When the direction of the microseisms was the same as that of the infrasound determined from our tripartite array (which was often the case), the amplitudes (power) were compared using the autocorrelation values. The I/M ratios determined for different times of the day are shown in Table 3 for winter. We note that the changing ratio in winter is in accord with the physical description of changing reflection height during the day in a dissipative atmosphere. The ratio for summer is greater, indicative of the lower level of reflection evident in Fig. 1. It is this type of analysis, applied on a continual basis, that allows the possibility of estimating reflection heights independently, once the "model" calibration is performed which specifies the height of reflection for a given I/M ratio.

4. Atmospheric tidal circulation in winter

Having determined the approximate height of reflection we are in a position to relate some statistical summaries of circulation patterns that predominate at 100–115 km.

a. Semidiurnal oscillation

In our latitude (40°N) the semidiurnal tide is quite effective in causing amplitude variations of infrasound in the manner previously described. As we record the signal continuously it is possible to give a meaningful representation of how often this tide provides for reflection from the 105 km region. Between October and mid-April the ocean is nearly always agitated enough to generate relatively strong infrasound as shown by the presence of microseisms; hence the prominent amplitude variations due to tidal effects ought to be observed. Thus, if we fail to see an amplitude oscillation with semidiurnal character this could mean one of two things: either the tide is absent or the prevailing wind and temperature have altered sufficiently to prevent tidal effects from causing reflection below the dissipation level. We cannot distinguish between these two effects but merely note the percentage of times the semidiurnal tide allows for the normal pattern of amplitude variation with east winds at 105 km about 1200 and 2400 LT. In winter (defined here as October to mid-April) analysis shows that this occurs 62% of the time. Thus the semidiurnal tide as modelled by Table 1 is a fairly regular feature of the winter circulation. Because prevailing wind and temperature changes presumably do occur, this percentage should be thought of as the lower limit of semidiurnal tide reliability.

Note that we have referred to the semidiurnal oscillation as being one of wind; incoherent scatter results (Fontanari and Alcayde, 1974; Bernard, 1974a) show that a semidiurnal oscillation in temperature also exists. Its amplitude is of the order of 20–30 K; in view of the expected tidal wind variation of 45 m s^{-1}, the square-root dependence of sound velocity on temperature indicates that the average temperature oscillation contributes only one-fourth that of the wind.

b. Diurnal oscillation

The existence of a diurnal oscillation in the upper winds in our vicinity is discernible in data already presented (Donn and Rind, 1972). According to Fig. 3 a different amplitude pattern is obvious in the early morning compared to the early evening; higher infrasound amplitudes at 0400 suggest east winds around 105 km. Because this is a summary of data the lower amplitude peak in Fig. 3 at 0400 compared with the two main peaks is indicative of the sporadic nature of the occurrence.

Careful scrutiny of Table 3 reveals that the directional rotation is not 12 h apart but is quicker in the morning than afternoon (120° azimuth at 0200 and 1200). Also listed in Table 3, is the percent of time that the predominant signal was from the west. This signal comes from sources in the Pacific Ocean and is quite common, being reflected by stratospheric westerlies far below any level of dissipation. It is the primary signal only when signal from the east is quite weak. According to Table 3 dominant signal from the west occurred more frequently from 1600 and 1800 LT when the east signal is weak (Fig. 3) than from 0400 and 0600. All this suggests a diurnal wind component with maximum east wind at 0400 added to the predominant semidiurnal tide.

As with the semidiurnal oscillations, continuous observation allows us to investigate the regularity of the diurnal effect. In the same context as the percentage given for the semidiurnal tide for 6 years of winter data, we see a diurnal oscillation with maximum east wind at approximately 0400 (and no corresponding effect at 1600) 42% of the time. Although the phase of this oscillation does vary, our data show that 0400–0600 is by far the most common time for maximum east winds. We conclude that the diurnal tidal wind oscillation must be of quite strong amplitude to produce east winds at 0400 in the same region that semidiurnal winds are east at 1000 and thus west at 0400.

Our observations of records during stratospheric warmings show that the diurnal tide (as well as the semidiurnal tide) has quite a strong amplitude at these times with maximum east winds invariably at 0400. This suggests that the thermospheric tide is coupled to the circulation patterns below.

5. Summary and conclusions

We have shown here and in previous papers that infrasound reception (from marine sources) is controlled by the behavior of the atmospheric tides. By inversion we use infrasound behavior to determine tidal circulation. We also show how height determinations can be made on a dependent basis with reference to an atmospheric model and on an "independent" basis with the use of infrasound/microseism ratios.

The advantages of using infrasound as a monitor of the upper atmosphere are numerous. In conjunction with a technique which measures either wind or temperature, the sound velocity determination would allow the other to be known. Its continuous operation capability, inexpensive nature, and on-line data processing present the possibility of practical everyday use to determine the regularity of the diurnal and semidiurnal tide. For example, at Palisades, N. Y. (41°N, 74°W), we have concluded that the semidiurnal tide in winter is present at least 62% of the time, and the diurnal tide at least 42% of the time. These results are based on daily observations in winter for 6 years.

It would be especially interesting to compare these results with a simultaneous recording of ionospheric drifts. The close concordance of the results noted in the literature [see Briggs and Spencer (1954) for ionospheric drifts] between the two techniques convince us they look at the same level. There is no doubt that the infrasonic technique observes the neutral atmosphere winds (and temperatures) and comparison would help validate both techniques.

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APPENDIX

TABLE A1. Survey of published wind and temperature observations. (References are identified in Table A2.)

Station	Location	Technique	Reference
Carnarvon, Australia	25°N, 114°E	rocket	1
Sonmiani, Pakistan	25°N, 66°E	chaff	2
Cape Kennedy, Fla.	28°N, 80°W	rocket	3, 4, 5
Eglin AFB	29°N, 87°W	chemical trails	6–18
Hammaguir, Algeria	31°N,	chemical trails	19
Yamagawa, Japan	31°N, 131°E	ionospheric drift (I.D.)	13, 20–25
Woomera, Australia	31°S, 137°E	rocket	26, 27
White Sands, N. M.	32°N, 106°W	rocket	4, 8, 11, 12, 28–32
Yuma, Ariz.	33°N, 114°W	chemical trails	15, 33–36
Pt. Mugu, Calif.	34°N, 119°W	rocket	5, 37
Adelaide, Australia	35°S, 139°E	meteor radar	38–50

APPENDIX (*continued*)

Station	Location	Technique	Reference
Washington, D. C.	37°N, 77°W	I.D.	51, 52
Ashabad, U.S.S.R.	37°N, 58°E	I.D.	53
Dushanbe, U.S.S.R.	37°N, 67°E	meteor radar	54
Arenosillo, Spain	37°N, 6°W	chaff	55–58
Tonopah Range, Nev.	38°N, 117°W	chaff	59
Wallops Island, Va.	38°N, 75°W	rocket, chemical trails	3, 5, 11, 13, 37, 60–71
Chiquita, Argentina	38°S, 57°W	rocket	5
Stamford, Calif.	38°N, 122°W	meteor radar	72
College Park, Pa.	40°N, 77°W	low frequency I.D.	73, 74
Sardinia, Italy	40°N, 90°E	chemical trails	71
Bedford, Mass.	42°N, 71°W	meteor radar	75, 76
Frunze, U.S.S.R.	42°N, 69°E	meteor radar	54, 77
Durham, N. H.	43°N, 70°W	meteor radar	72
Millstone Hill, Mass.	43°N, 71°W	incoherent scatter	78–80
Birdlings Flat, New Zealand	44°S, 173°E	partial reflection I.D.	81
St. Santin, France	45°N, 2°E	incoherent scatter	82–90
Ottawa, Canada	46°N, 76°W	I.D.	21, 52, 91
Garchy, France	47°N, 3°E	meteor radar, I.D.	82, 92–97
Rostov-Don, U.S.S.R.	47°N, 39°E	I.D.	53, 98
Freiburg, Switzerland	47°N, 7°E	I.D.	21, 24, 91, 99, 100
Volgagrad, U.S.S.R.	49°N, 45°E	rocket, chemical trails	101, 102
Domont, France	49°N, 2°E	I.D.	91
Kharhov, U.S.S.R.	50°N, 33°E	meteor radar	53, 103
Collin, G. D. R.	51°N, 13°E	low frequency I.D.	104–110
Cambridge, Gr. Br.	52°N, 0°	I.D.	52, 111–113
Swansea, Gr. Br.	52°N, 4°W	I.D.	112, 114, 115
Saskatoon, Canada	52°N, 107°W	partial reflection I.D.	116–118
Debilt, Neth.	52°N, 5°E	I.D.	22, 119–121
Irkutsk, U.S.S.R.	52°N, 104°E	I.D.	53, 122–124
Sheffield, Jodrell Bank	53°N, 2°W	meteor radar	41, 42, 45, 48, 75, 125–128
Kuhlungsborn, G. D. R.	54°N, 12°E	low frequency	104–110
Primrose Lake, Canada	50°N, 110°W	rocket	5
Obninsk, U.S.S.R.	55°N, 35°E	meteor radar	129
Models or data compilations from latitudes 25° to 55°			11, 13, 45, 46, 53, 54, 67, 101, 114, 128, 130–154

TABLE A2. References for Table A1 citations.

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