A Climatological Study of Strong Downslope Winds in the Boulder Area

Waltraud A. R. Brinkmann

NCAR Cooperative Thesis No. 27
INSTAAR Occasional Paper No. 7

University of Colorado and the
Laboratory of Atmospheric Science
and GARP Task Group, NCAR

1973

OFFICE COPY
DO NOT REMOVE FROM THIS FILE

The National Center for Atmospheric Research (NCAR) is operated by the nonprofit University Corporation for Atmospheric Research (UCAR) under the sponsorship of the National Science Foundation.

A Climatological Study of Strong Downslope Winds in the Boulder Area

Waltraud A. R. Brinkmann

A thesis submitted to the Faculty of the Graduate School of the University of Colorado in partial fulfillment of the requirements for the degree Doctor of Philosophy, Department of Geography, based on research conducted in cooperation with the scientific staff of the Laboratory of Atmospheric Science and the GARP Task Group of the National Center for Atmospheric Research, Boulder, Colorado.

Miss Brinkmann received her B.S. and M.S. degrees from the University of Calgary in 1967 and 1969, respectively.

NCAR Cooperative Thesis No. 27
INSTAAR Occasional Paper No. 7

University of Colorado and the
Laboratory of Atmospheric Science
and GARP Task Group, NCAR

1973

ABSTRACT

No systematic study of the characteristics of Boulder's severe wind storms has ever been made because of lack of adequate data, but their complexity is known from a few successful attempts of flying into them and some theoretical work. Instruments operated by a number of agencies are now providing a set of surface data covering several years. This permitted the selection of 20 severe storms, with gusts exceeding hurricane-force speeds, for a climatological study.

Boulder's storms have generally been considered to be part of the system of warm Chinook winds occurring in the lee of the Rocky Mountains (these are generally defined in terms of marked surface temperature rises at stations at the foot of the mountains) but of the 20 storm cases almost half were found to have been genetically cold although most of them appear as warm winds at lower levels because of the removal of surface inversions or supression of their development and diabatic warming. However, those storms that arrive still cold at the foot of the mountains produce temperature decreases equivalent to those observed with the classic Bora of Yugoslavia. The warmth of the genetically warm storms in Boulder was not found to be the result of the classic 'foehn effect' because of little, if any, precipitation on the windward slopes.

The storms were found to occur most frequently in the morning hours in the month of January. Their gustiness was found to be much higher than averages determined for the temperate zone but similar to that of other warm and cold downslope winds around the world. Many other characteristics of Boulder's storms, such as the narrow belt affected by high winds, the interruptions or pauses, fluctuations in surface pressure, are features common to downslope winds in other parts of the world. It is hypothesized that all downslope wind storms may be generated by a similar mechanism.

Conditions favourable for Boulder's wind storms were found to be the existence of a stable layer above mountain top as well as relatively strong winds at that level but speeds not too high in the upper troposphere. It appears that these air mass characteristics lead to the development of a deep 'lee flow disturbance' (resembling a lee trough or hydraulic jump) over the eastern slopes. Air is accelerated down the slopes toward the local pressure minimum beneath the disturbance but is decelerated again rapidly to the east. At the height of the storms in Boulder, mean hourly speeds are on the average 20 m s^{-1} with mean maximum gusts of 36 m s^{-1}, at the same time speeds along the lee slopes to the west and within 10 km to the east into the plains are 50 per cent less. Changes in the critical air flow characteristics upwind result in shifts of the 'lee flow disturbance' and, consequently, movements of the surface wind maximum along the slopes. Thus, wind maxima on the slopes and out into the plains do not tend to coincide with storms in Boulder.

The results of this study should present a basis for further work, such as observational and statistical forecasting, modelling of the storms, and more detailed analyses of individual storm cases.

ACKNOWLEDGMENTS

It is with pleasure and deep gratitude that the author takes this opportunity to thank the many people and organizations which contributed to the attainment of this goal. Foremost among these are Dr. R. G. Barry, who gave much helpful guidance and encouragement, and Dr. E. J. Zipser, who first suggested the occurrence of cold storms in Boulder and whose assistance has been invaluable.

The author wishes to thank the following individuals and organizations:

Drs. D. K. Lilly and J. B. Klemp for many helpful discussions and critical comments;

Dr. H. W. Baynton for his help with some of the technical aspects of the wind data and the setting up of instruments;

Drs. P. R. Julian and R. Bleck for providing special computer programs;

Dr. H. F. Quick and the Manager of the Boulder Airport for permitting the setting up of weather stations on their property;

University of Colorado's Institute of Arctic and Alpine Research for providing part of the basic data and additional instrumentation;

National Center for Atmospheric Research (which is sponsored by the National Science Foundation) for providing part of the basic data, additional instrumentation and computer time;

United States Forest Service, Rocky Mountain Forest and Range Experiment Station, Fort Collins, for the Mines Peak data;

United States Weather Service Forecast Office, Denver, for the Denver data;

Colorado State Climatologist for the Eagle data;

University of Colorado for fellowships covering three winters.

Mr. W. Wilson digitized some of the wind data; the cartography was done by Ms. K. Verhulst, the photography by Mr. R. Carter and the typing by Ms. G. Ericksen.

Finally, the author greatly appreciated the opportunity to participate in the NCAR lee wave flight program -- a memorable experience.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

CHAPTER I

BACKGROUND TO THE STUDY

1. Introduction

The principal feature of Colorado's geography is its position astride the highest mountains of the Continental Divide. Boulder is located at the transition between plains and mountains. The latter rise abruptly at the western edge of the city: elevations increase by about 1500 m between Boulder and the Divide, about 30 km to the west. The general topography and climatology of the area have been discussed by Paddock (1964) and Barry (1972).

An outstanding feature of Boulder's winter weather is the severe downslope wind storm which has generally been considered to be part of the system of warm Chinook winds occurring in the lee of the Rocky Mountains from Alberta/Canada to northeastern New Mexico/U.S.A. (for example, Harrison 1956; Beran 1967; Houghton and Kasahara 1968; Julian and Julian 1969; Sangster 1970; Vergeiner and Lilly 1970; Vergeiner 1971). 'Chinook' is the local name of the class of warm winds which occur in the lee of many mountain ranges around the world and are known under the generic name 'foehn'. The counterpart of the foehn is the cold bora. Both are generally identified with respect to temperature conditions because of the sudden and marked changes in surface temperature (rises/drops) found to be associated with them (discussed further in Section 2).

To the people living in the Boulder area temperature changes associated with the downslope wind storms do not seem to be remarkable, however. They also appear to be less familiar with the term Chinook than residents of the northern part of the Chinook belt. The local cartoonist for the Calgary (Alberta/Canada) newspaper, for instance, never misses a winter without at least one of his creations dealing with the Chinook phenomenon. The apparent lesser importance, to the residents of the Boulder area, of temperature changes associated with the wind storms are partly due to Boulder's location at a lower latitude as well as at the foot of the mountains where wind speeds are highest (as will be shown in Chapter III). Furthermore, because of the north-south trend of the Rocky Mountains, cold as well as warm air masses can cross the mountains from west to east and descend the lee slopes in Colorado. It might therefore be reasonable to suspect that some of Boulder's wind storms exhibit bora-type characteristics. As a matter of fact, 'old timers' of the Boulder area must have already been aware of unorthodox decreases in temperature associated with some storms as the heading of an article in the "Boulder Daily Camera" on 26 January 1910 suggests "OLD BOREAS BLEW HIS CHILLY BREATH - Boulder in the teeth of gale of great velocity which whistled through streets, creating some havoc" (Fig. 1.1). Besides the interesting comment regarding the coldness of the wind and the use of the appropriate generic name, the article points out a few of the typical characteristics of Boulder's storms which will be discussed in Chapter III, such as the nighttime occurrence and the little damage done in the mountains compared with that occurring at the foot of the Front Ranges.

People in the Boulder area are more concerned with the occasional extremely strong and violent downslope winds or 'wind storms'. Gusts, sometimes exceeding 45 m s^{-1}, have overturned mobile homes, trucks and aircrafts, blown trains off the track and unroofed houses. They have blown down trees, power poles and fences, and found to be

OLD BOREAS BLEW HIS CHILLY BREATH

BOULDER IN THE TEETH OF GALE OF GREAT VELOCITY WHICH WHISTLED THROUGH STREETS, CREATING SOME HAVOC

People with uneasy consciences didn't sleep a wink last night, for it blew a gale. Electric wires made merry music and tin signs started a terrible diapason. Street lights went out and for only a brief period homes were in darkness owing to crossed or fallen wires. Some citizens who should have easy consciences also complain of insomnia. Streets were whipped dry of what little snow remained and the dust went rolling Valmontward.

The roof of the Curran Opera house was torn out, resulting in a damage of about $250. Manager Penney went to Denver this afternoon to make arrangements to have a new one put on before tomorrow night's production of the Primrose Minstrels. The opera house will be dark tonight. The company suffered considerable loss from the blowing down of billboards and other signs. W. T. Story has agreed to have the roof replaced by tomorrow afternoon.

The large chimney of the McAllister Lumber company was blown down, as were several electric light posts on 15th street. The glass front of the German house, on West Pearl street, was blown in, as were also those of T. S. Waltemeyer's office and Chas. Ingram's bicycle shop on 13th street.

The roof over the back porch of the large storeroom and house at 1713-1715 15th street, was torn off and thrown into the middle of the street. The wind picked it up and carried it over the rest of the house and dropped it into the street in two big sections. The house is occupied by Mrs. Merrill.

Several oil derricks were blown down in the oil belt, and considerable damage done to barns and farm houses east of this city.

The University was damaged but slightly. Windows were blown in at the University hospital; tiles were blown off the Library, and part of the iron roof was torn off the Engineering building. The power plant, which was damaged about $2,500 a short time ago, was uninjured by last night's winds.

Window lights were blown in all over the city, trees and telephone posts were torn down, and the wind played havoc with awning and sign boards on Pearl street.

The wind did considerable damage at Marshall, but there was no fire there, as reported on the streets this morning. A prairie fire east of Marshall was probably responsible for the rumor. Fires were also reported from Broomfield and Gunbarrel Hill, but this has not been verified.

The wind did little damage in the mountains except the putting out of commission of the city's telephone line to Silver Lake. The Denver, Boulder & Western railroad reported 4 inches of snow at Cardinal.

The school at White Rock was dismissed today, the board not considering the building safe owing to the wind, which caused a large split in the west wall. Immediate repairs are contemplated.

During the heavy wind last night the prairie adjacent to and east of Burns Junction caught on fire and was burning fiercely at midnight. A number of haystacks in the path of the flames had been consumed and others were burning, which, together with the rapidly expanding fire fanned by the gale, presented an unusual spectacle to passengers on the late Interurban trains.

The large chimney of the Jefferson school house was blown down last night and crashed through the roof into the 8th grade room, smashing some furniture.

One Colorado Oil & Refinery company, two Boulder-Greeley and the one remaining Otero oil derrick, were blown over during the progress of the windstorm.

The wind blew down the porch of Mrs. Frank Grigsby, at 940 15th street, and the large plate glass window was also blown in.

The large wagon of the Imperial Tea & Coffee house was blown over near 12th street bridge last evening, and Harvey Pence, the drive, considerably bruised. Dishes, a case of eggs and other produce were smashed to pieces.

Fig. 1.1. A newspaper article that appeared in the Boulder Daily Camera on 26 January 1910.

responsible for fires and for dust storms serious enough to impede traffic (Julian and Julian 1969). The three most severe storms over the last 6 years have caused property damage estimated at $0.75 million, $1.5 million and $2.5 million respectively and together injured over 50 people and killed two. During the 7 January 1969 wind storm 50 per cent of the residences received damage (City of Boulder 1970).

Adjustments to the hazard are, however, sadly lacking. Because of the beautiful view to the mountains, most residences have their picture windows facing west; consequently, 24 per cent of a sample of Boulder residences received damage to windows during the 7 January 1969 storm since less than 4 per cent had window shutters installed (City of Boulder 1970). Similarly, in a badly damaged mobile home park over 50 per cent of the trailers were found to have had their long axis oriented north-south, and the damage to these accounted for over 70 per cent of the damage to the whole park and over 80 per cent of the total destruction (Miller 1972). On the other hand, the old central building of a farm house in this area (dated to pre 1893 because of the use of square nails in the construction which was discontinued at that time) has no windows to the north or west. Additional rooms have been added on to it since, including a room with a window facing west (Grossman 1973, pers.comm.).

Boulder is a very fast growing city; its population tripled over the last 20 years; consequently, there are many new residents in the area who are relatively unfamiliar with the infrequently occurring very severe storms. Miller (1972) also found that people tend to underestimate the severity of the hazard, particularly in high income groups. Unfortunately, it appears that action from the local government is precipitated only by the occurrence of very severe wind storms. After the 7 January 1969 storm, for instance, restrictions were placed on roof composition and fencing technique and the 11 January 1972 storm prompted the requirement for mobile home tie downs and some improvements in the building code (Miller 1972).

From the above discussion it is apparent that the high velocities of Boulder's winds are much more important than any temperature changes associated with them. The complex nature of these severe storms is known from casual observations, a few successful attempts of flying into them and some theoretical work. However, except for newspaper surveys and examinations of the U.S. Weather Bureau Storm Damage Records as far back as 1905 by Julian and Julian (1969) and back to 1869 by Whiteman and Whiteman (1973, pers.comm.) no work has been done on the climatology of the storms because of lack of adequate wind data. Instruments operated in the Colorado Front Range-Boulder Valley area by a number of agencies are now providing a set of surface data covering several years. This permitted the selection of 20 severe wind storm cases, with gusts exceeding hurricane-force speeds, for the present study. At this stage of knowledge about Boulder's wind storms the climatological approach was considered to be the most fruitful.

The purpose of this study was therefore to produce a climatology of the storms upon which further work may be based, such as statistical and observational forecasting, the modelling of the storms, and more detailed studies of individual cases. An attempt was also made to present the results in a global perspective by comparing the characteristics of Boulder's wind storms with those of other winds around the world. This was somewhat difficult because, as mentioned above, the emphasis has been more on the temperature characteristics as compared with the associated surface wind speeds. Particularly in studies of foehn-type winds, high surface velocities may or may not have accompanied the cases analysed.

2. Global Summary of Foehn and Bora Winds

a. Foehn Winds. 'Foehn' was the name originally given to warm, dry winds descending the northern slopes of the European Alps. Due to the early extensive work in the Alpine region, the term 'foehn' has become so general that it is now the generic name for all warm, dry downslope winds, although they are also known by a variety of local names such as the Chinook of the Rocky Mountains, the Zonda and Puelche of the Andes, the Nor'-wester of New Zealand, the Santa Ana of Southern California, the Berg Wind of South Africa, the Kachchan of Ceylon, the Afganet and Ibe of Central Asia, the Germich of the Caspian Sea, etc.

The outstanding characteristics of foehn-type winds are generally considered to be their warmth and dryness which is reflected in the definitions used in many studies to select foehn wind cases (for details see Brinkmann 1971). According to the most commonly used definition the wind is characterized by a simultaneous and abrupt increase in surface temperature and decrease in humidity (together with a surface wind from the direction of the mountains).

The emphasis on warmth, which has led to the application of the term to almost any sudden and outstanding rises in temperature observed in the lee of mountain ranges (Glenn 1961), is understandable considering the saturation deficit associated with large rises in temperature and the accompanying low humidity. A survey of the great amount of literature dealing with the statistics of those warm winds (Brinkmann 1969a) showed that maximum temperature rises are on the average 10°C for the European Alps and 15°C for the Rocky Mountains, although extreme cases of a rise exceeding 20°C have been reported (Conrad 1936; Johnson 1936; Hader 1946; Ives 1950; Turner 1966). The relative humidity associated with the wind in both areas is usually below 50 per cent and frequently drops to 20 per cent. There can therefore be rapid increases in melting, evaporation and sublimation. Foehn winds have been reported to remove snow at a rate of one inch per hour which is welcome to the cattlemen of the High Plains because of the clearing of pastures during the winter season (Marsh 1965), but the removal of the snow as well as the drying of the soil can also have serious effects on the moisture budget in semi-arid lee areas and can lead to subsequent soil erosion (Ashwell and Marsh 1967; Ashwell 1971). Increased transpiration during the cold season when the ground is frozen and rapid changes in temperature are known to be injurious to vegetation (Henson 1952; MacHattie 1963). Accompanied by high velocities, foehn-type winds are also known to set the stage for serious fires (Sergius _et al._ 1962; Virgo 1966).

Since the foehn has usually been considered to be a temperature phenomenon, cases with strong surface winds may have been included in foehn studies but few separate detailed analyses of the occasional foehn storm exist. The lesser importance placed on surface wind speed is also reflected in the more common use of wind direction rather than wind speed as a criterion for foehn. Consequently, surface speeds associated with warm winds have been found to vary from almost calm to gale strength. Strong, gusty wind and destructive violence have, however, been reported from many foehn areas (Schmitt 1930; Mirrlees 1934; Georgii 1954; Sergius et al. 1962; Aanensen 1964; Borisov 1965; Bluethgen 1966; Virgo 1966; Rumney 1968; Arakawa 1969). The Zonda of the Andes, for instance, has been reported to reach 35 to 40 m s^{-1}, dislodging stones in its path. The Santa Ana of Southern California, with gusts up to 40 m s^{-1}, has unroofed and flattened houses. The Germich of the Caspian Sea often reaches speeds of 20 to 25 m s^{-1} which seriously affects sowing as well as the harvesting of fruit. A maximum of 58 m s^{-1} was recorded at Kangerdlua fjord, East Greenland, during the British Arctic Expedition 1930-1931 before the instrument blew away. Destructive gales during the Canterbury Nor'-wester of New Zealand have raised clouds of dust from river beds. During a particularly strong and gusty wind storm in the lee of the Pennines of England (Fig. 2.1) which resulted in extensive damage to houses and woodlands, the forester said that he could smell burning "caused by trees rubbing one against another" (Aanensen 1964, p.12). On Hokkaido, Japan, warm downslope winds with gusts exceeding 45 m s^{-1} (Fig. 2.2) completely destroyed wooden houses and caused severe damage to forest. The strength and gustiness of the Foehn of the northern slopes of the European Alps (Fig. 2.3) is suggested by reports of uprooted trees

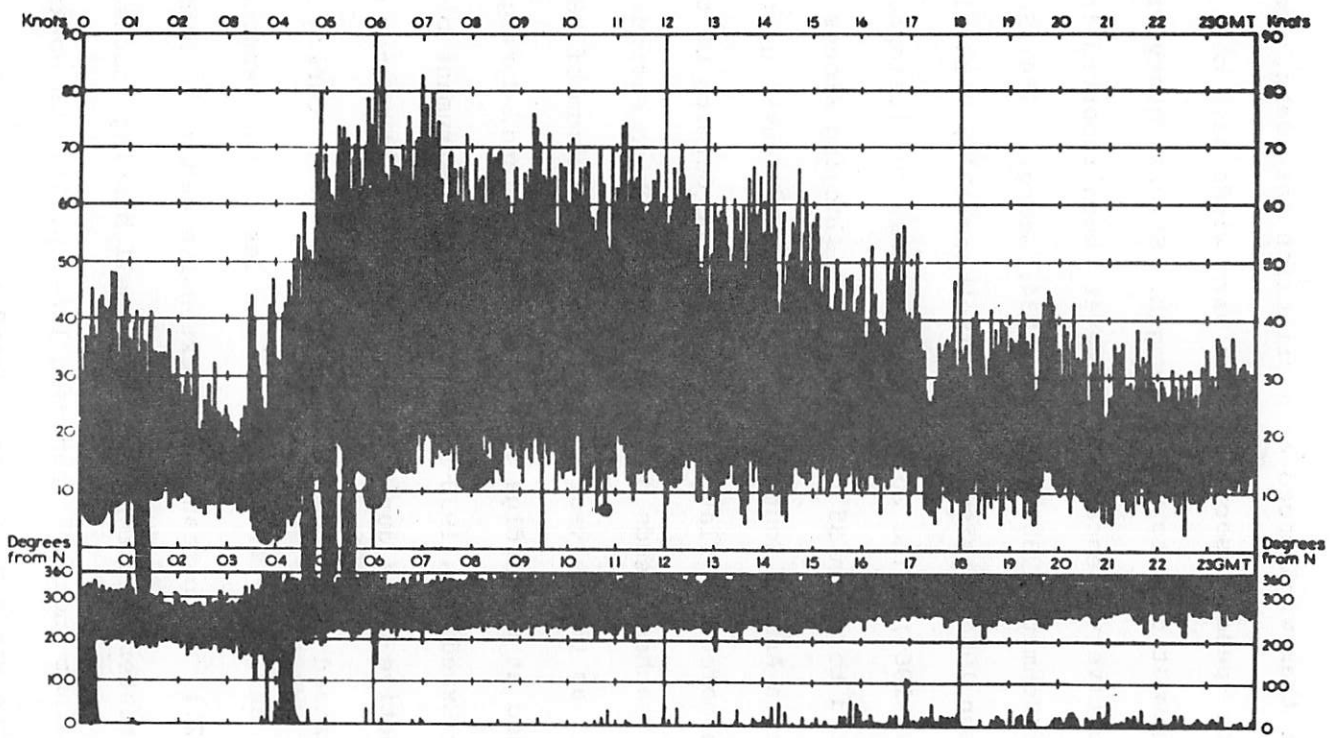


Fig. 2.1. Anemometer trace from Sheffield/England on 16 February 1962 (from Aanensen 1964).

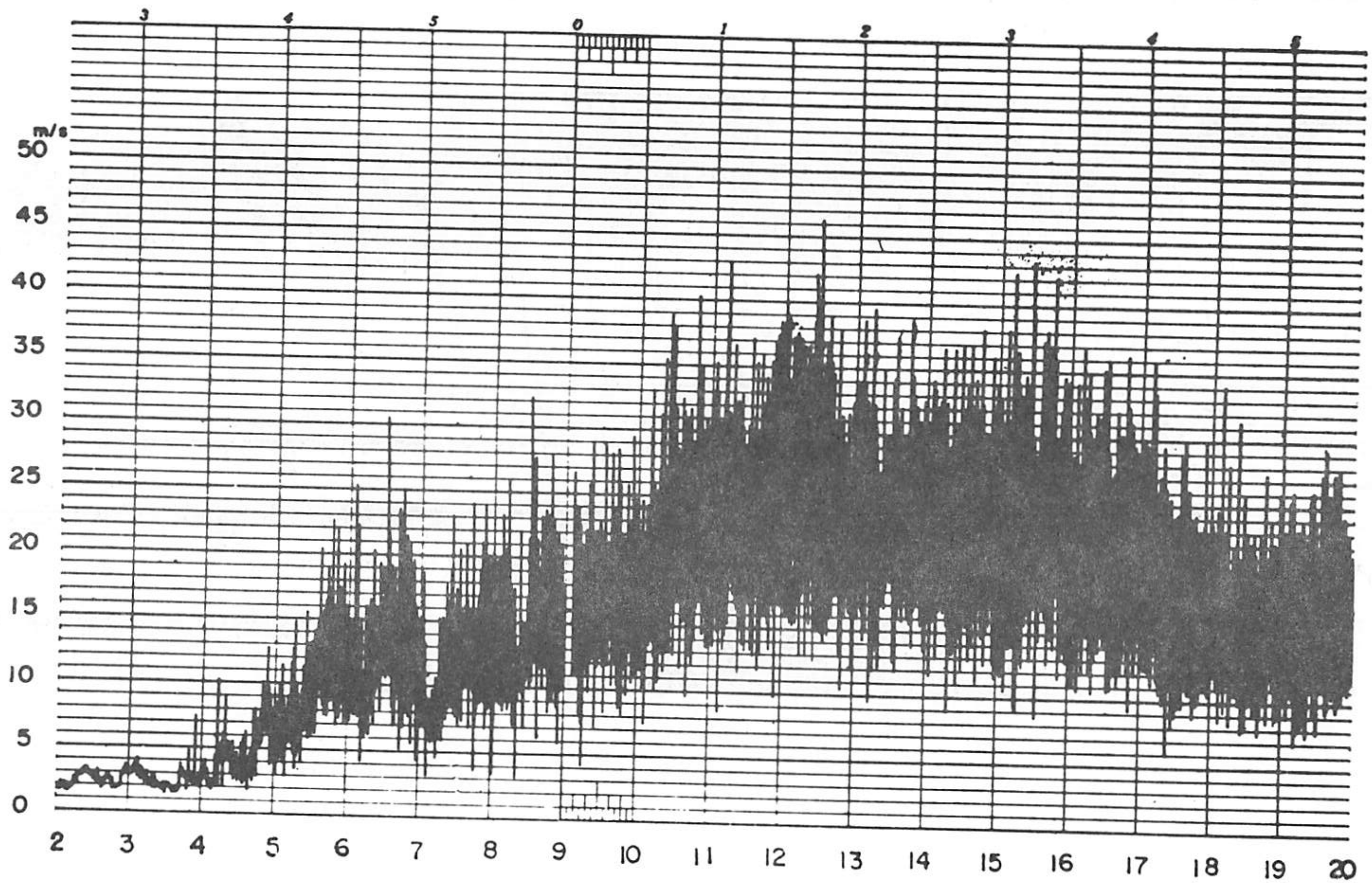


Fig. 2.2. Anemometer trace from Hokkaido/Japan on 27 September 1958 (from Arakawa 1969).

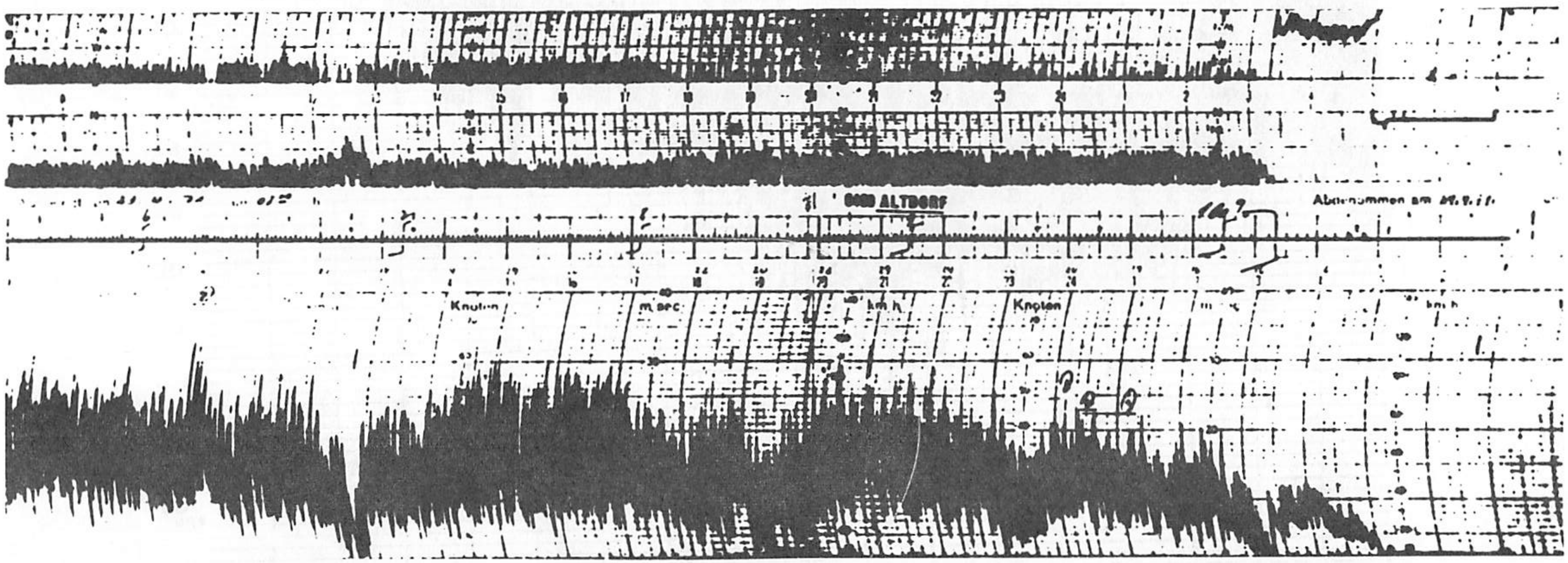


Fig. 2.3. Anemometer trace from Altdorf/Switzerland on 23 April 1971 (from Gutermann, pers. comm.).

and broken telephone poles and by the practice of placing heavy rocks on roofs to prevent them from blowing off. Strong gusty foehn-type winds are also known to endanger shipping on the larger lakes of the Alps and along the south shore of the Caspian Sea.

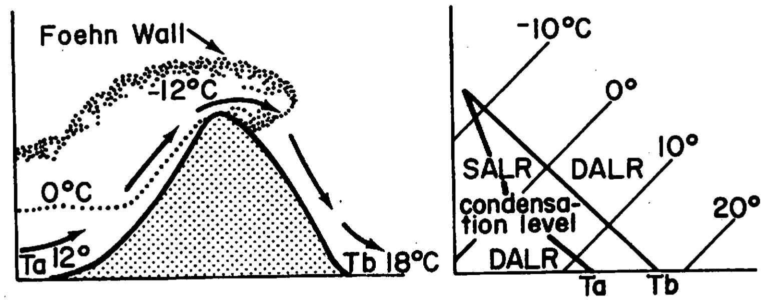
The width of the belt affected by such strong winds has been found to be much smaller than that affected by the temperature rise accompanying foehn winds. The violence of the Foehn, for instance, decreases rapidly with increasing distance north of the Alps and beyond about 60 km the Foehn is noticeable only by a small increase in temperature and decrease in relative humidity, very seldom in terms of air movement (Schmitt 1930), while the temperature effect may sometime be felt beyond about 100 km to the north. On Hokkaido, Japan, the area of damage was found to extend not much beyond 10 km from the foot of the mountains. East of the Rocky Mountains the Chinook belt, in terms of temperature effects, is estimated to be about 300 km wide in Canada (Longley 1967), seems to get narrower toward the south and in Colorado does not appear to extend much beyond 100 km (Lovill 1969); the width of the belt of high winds has, however, not yet been determined.

A frequently mentioned characteristic of warm downslope winds is the 'foehn pause' which is a temporary cessation of the foehn at the ground in terms of a rapid fall in temperature and increase in relative humidity and/or a decrease in wind speed. Another feature commented upon by several authors is the fluctuation of surface pressure. Hann considered "local depressions", which he found in the Foehn valleys, to be the cause of the "extraordinary local violence of the stormy foehn winds" (Hann 1903, p.354).

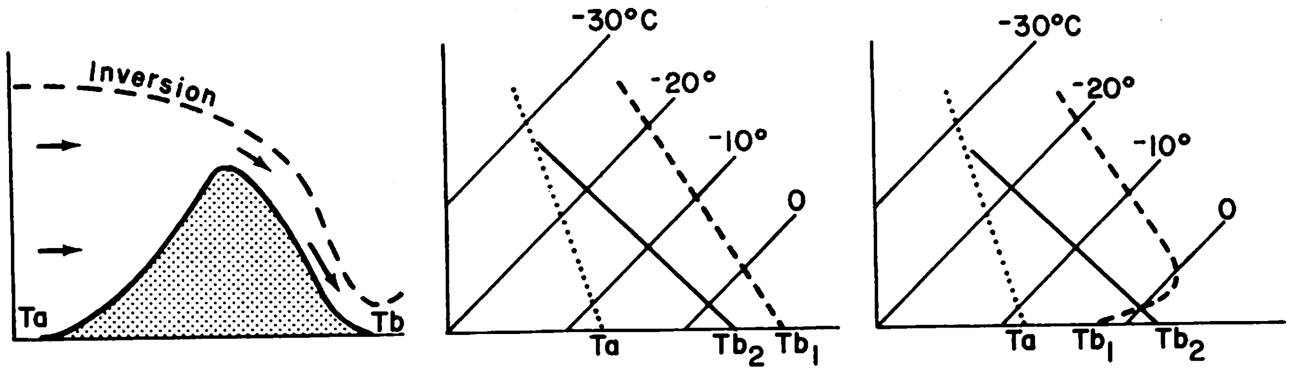
Mirrlees (1934) found the advances and retreats of the East Greenland gales between the Ice Cap station and the fjord to be accompanied by irregular changes in pressure with a sudden decrease of 5 to 10 mb at the time of arrival; he did not observe such pressure changes during gales from other directions.

The mechanism which has been considered to be the major cause of the warmth and dryness of descending air had already been identified by the American meteorologist Espy as early as 1835 but credit for the foehn theory generally goes to the Austrian Hann. In 1866, Hann showed that as air rises on the windward slopes and cools, moisture is condensed and removed from it, and that the latent heat liberated during this process causes the temperature of this air to be higher, altitude for altitude, in the lee of the mountain range than on its windward side (Fig. 2.4). Moreover, he was able to prove his point by describing a similar wind in Greenland which could not possibly be due to transport of air from a warmer region (according to an older foehn theory). Later Hann (1885) added that precipitation does not always have to occur on the windward side since adiabatic compression of descending air will result in relative warmth and dryness. This has recently been demonstrated by Cook and Topil (1952) and Lockwood (1962) but precipitation on the windward side is still considered to be the chief characteristic of the foehn process (Ives 1950; Defant 1951).

Although the general thermodynamic explanation of the warm, dry foehn winds was already known more than 100 years ago, the question of the mechanism causing the warm air to descend and displace the colder, denser air in the lee has not yet been answered



The foehn effect. Ta refers to the temperature at the windward foot of the range and Tb to that at the leeward foot (from Barry and Chorley 1968).



The bora effect (center) and cold air foehn effect (right). Ta refers to the temperature sounding for a windward station, Tb_1 refers to the sounding for a station in the lee prior to the onset of the bora, Tb_2 represents the descent of a parcel of cold air originating on the windward side.

Fig. 2.4. The thermodynamics of the foehn, bora and cold air foehn effects.

to everybody's satisfaction. Over the years a number of theories have been advanced to explain the process, such as the solenoid field developing in the lee to give the air a downward velocity component (Frey 1953), the frictional drag of the mountain barrier on the upper air current (Wild 1901), the withdrawal of cold surface air which necessitates replacement from above (Billwiller 1899), hydraulic jump and lee wave theory (for details see Brinkmann 1969b and Section 3).

Analysis of the pressure distribution during foehn with the aim of finding a unique synoptic pattern which might point to a possible driving mechanism of the wind has also not been very successful. In general, necessary conditions are considered to be a ridge of high pressure at the surface to the windward side of the mountain range and a trough to the lee, with a strong component of flow normal to the mountain range at upper levels. More detailed studies have however shown that there is no definite correlation between the foehn and any of the Grosswetterlagen (Hoinkes 1953; Obenland 1956). On the average there is a pressure gradient of 4 to 5 mb per 100 km across the Alps during Foehn (Frey 1957). Hoinkes (1950) however describes a case in which at the onset of the Foehn the surface pressure gradient across the Alps even pointed in the wrong direction (i.e., decreased from north to south). Such results are not surprising considering the problems associated with the definition of the foehn. Furthermore, in European literature two different foehn phases are recognized: the cyclonic or classical and the anticyclonic foehn. Warming associated with the latter phase, which is considered to be the precursor of the cyclonic foehn, is the result of subsidence in an anticyclone over the Alpine region; surface winds during this phase are light.

b. Bora Winds. The foehn wind's counterpart is the bora, a cold downslope wind which involves the movement of air masses, thus differing from the drainage wind which is the result of radiation cooling of a shallow layer of air and is seldom strong. Named after the best-known example of these cold winds, which blows down the western slopes of the Dalmatian Mountains on the west coast of the Adriatic, they are also known by a variety of local names such as the Elvegust and Sno of Norwey, the Mistral of southern France, the Bora of Novorossiisk (Black Sea) and Novaya Zemlya (Russian Artic), the N'aschi of the Persian Gulf, the Vardarac of the northern Aegean, the Helm Wind of the northern Pennines, etc.

One of the most outstanding characteristics of the bora is the sudden decrease in temperature which has a noticeable effect on the vegetation of the region, particularly on frost-sensitive plants such as fruit trees. The Mistral, for instance, is known for the frost damage it occasionally inflicts upon the vineyards of southern France. Maximum temperature changes associated with the Bora of the Dalmatian Mountains exceed on the average -10°C while the relative humidity can decrease to 20 per cent (Band 1951; Yoshino 1971). Vegetation in the bora areas tends to be sparse as well as dwarfed.

The violence and gustiness of bora winds (Fig. 2.5) is frequently emphasized in the literature by descriptions of heavy damage on land and endangered shipping on adjacent water bodies; windows and door

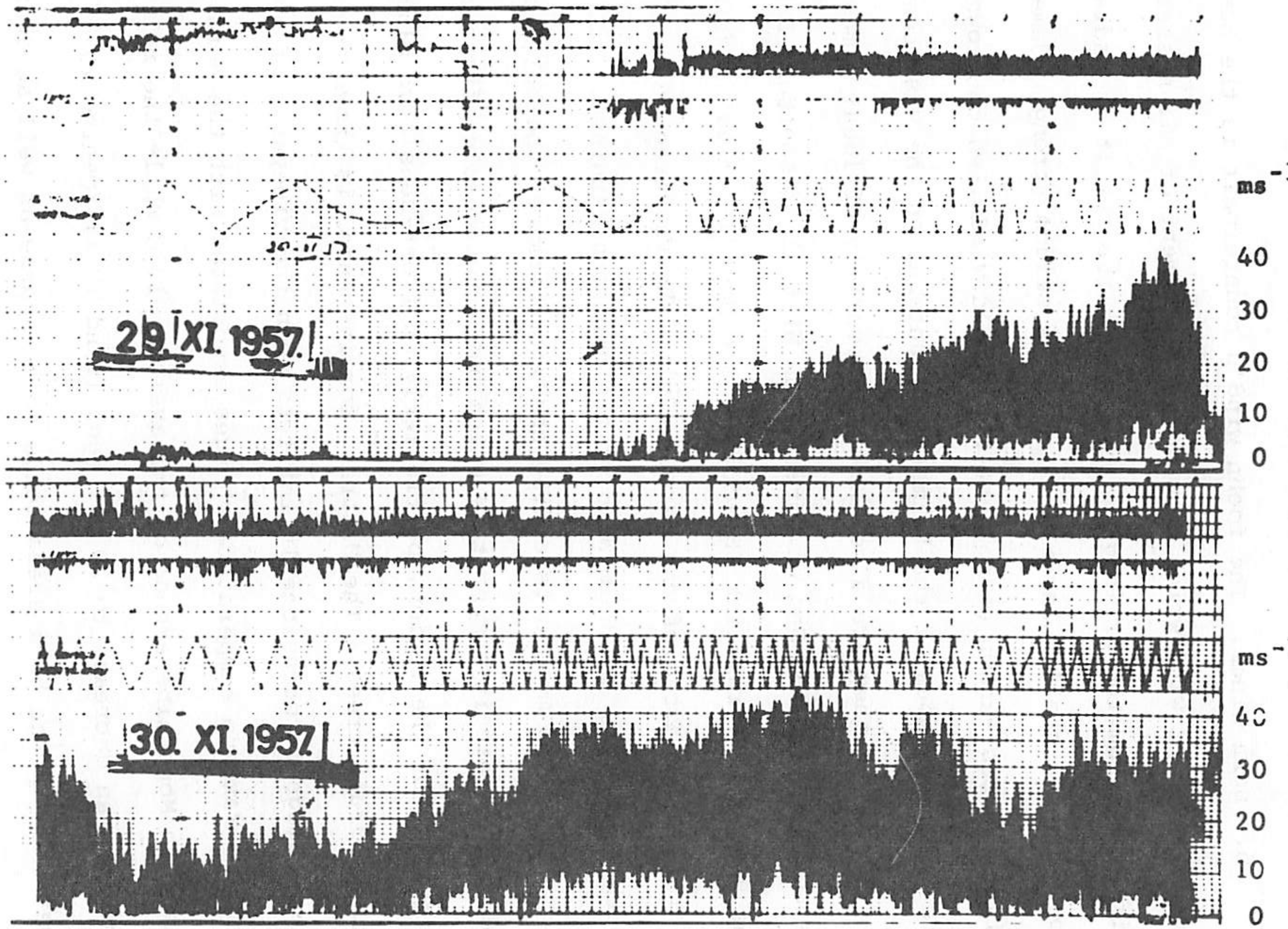


Fig. 2.5. Anemometer trace from Dubrovnik/Yugoslavia on 30 November 1957 (from Cadez, pers. comm.).

openings facing the mountains are usually kept to a minimum (Eckert and Klinkow 1947; Band 1951; Grober 1961; Borisov 1965; Bluethgen 1966; Gentilli and Fairbridge 1967). At Trieste, for instance, gust speeds of over 30 m s^{-1} are quite frequent and occasionally exceed 50 to 60 m s^{-1}. The wind is known to have lifted roofs, swept people off their feet and stopped traffic because of the many accidents caused by overturned automobiles, street cars and even trains; ropes are stretched along the sidewalks for use by pedestrians. During the Mistral wind speeds can exceed 35 m s^{-1} and at Novaya Zemlya velocities of 40 to 50 m s^{-1} have been recorded. The effect of these violent winds does not extend far beyond the foot of the mountains, however, but tends to decrease sharply within a few kilometers (Defant 1951; Ball 1956; Grober 1961).

Other striking features of bora-type winds are the occurrences of 'lulls' during which strong winds may be both audible and rendered visible by drifting snow higher up the slopes, and strong pressure fluctuations of several millibars associated with the onset and cessation of the wind (Loewe 1950; Band 1951; Ball 1956).

The dynamic principle behind the driving mechanism of the bora winds has never presented as much of a problem as that of the foehn. The bora is defined as a descending wind whose source is so cold that in spite of adiabatic warming the air remains colder, level for level, than the air that is being displaced and thus arrives at the foot of the mountain range as a cold wind (Fig. 2.4). It is this greater density which is assumed to be the cause of the violent descent since the colder air mass will experience a downward force proportional to the difference in density between it and the ambient air at each level. Comparisons between wind records from a mountain pass in the Dalmation Mountains and the Adriatic coast, for instance, showed no relationship and thus indicated that the Bora receives most of its momentum during descent (Band 1951). Manley (1945) also found that during the Helm Wind velocities at the crest of the Pennines are less than at their foot.

Synoptic conditions at the surface during the bora are somewhat similar to those generally considered to be necessary for the development of the foehn. There is usually a high pressure system on the windward side but a depression to the lee may or may not be developed. Thus, a distinction is made between anticyclonic and cyclonic bora, depending on whether the high pressure system or the low is the predominant synoptic feature. Band (1951) found surface wind speeds to be generally higher under the combined influence of a well developed high and low. At upper levels (500 mb) there may or may not be a component of wind normal to the mountain range (Band 1951; Yoshino 1971).

c. Mixed-type Downslope Winds. A major problem in the definition of foehn-type winds lies in the use of surface temperature rises observed at the foot of the mountains (where most people live). Temperature changes in the free air or even at the upper slopes are unknown or not considered, but are assumed to be similar to those at lower levels. This assumption is not always true. A cold downslope wind may, for instance, remove a strong surface inversion at the foot of the mountains (with potential temperatures lower than those of the descending cold air mass). The arrival of the cold air would thus produce a temperature rise at lower levels and result in an apparent foehn, or 'cold air foehn' (Fig. 2.4).

For example, east of the South American Andes, at latitude 33°S and elevation 750 m the city of Mendoza is in a location very similar to that of Boulder. It is situated at the bottom of the foothills which reach an average elevation of 3000 m about 75 km to the west and behind them, about 120 km west of Mendoza, the Cordillera rise to over 5000 m. The Zonda of Mendoza is considered to be a foehn-type wind producing temperature rises which may exceed 10°C and can be accompanied by high wind speeds. From discussions of surface conditions and some sailplane data (Georgii 1952, 1954, 1967) it appears, however, that the Zonda is genetically a cold wind. According to Georgii the Zonda consists of two phases covering a period of several days. During the first or 'stable phase' maritime tropical air crosses the Andes and descends in the lee but does not reach the surface at the foot of the mountains because of a shallow layer of cold air. Lenticular clouds can be observed during this phase. During the second or 'unstable phase' maritime polar air crosses the Andes from the southwest and descends the leeward slopes like a waterfall, having a cooling effect at the higher slopes but reaching Mendoza as a warm, strong wind. The Zonda is characteristic of the cold season, especially the month of August; it does sometimes occur in the summer but is then a cold wind in Mendoza.

Similar cold air foehns have been reported to occur at the foot of the south slopes of the European Alps and other mountain ranges (Ficker and Rudder 1948; Band 1951; Defant 1951; Cadez 1953; Georgii 1967; Reiter 1969). Even the Foehn of the northern slopes of the Alps has been suggested to have bora-characteristics at the upper slopes sometimes, and certainly the frequent discussions of the driving mechanism of the Foehn, in terms of removal of a cold layer at the surface, points in that direction (Kuettner 1940; Rossmann 1950; Frey 1953; Bluethgen 1966). The Santa Ana of Southern California may be another example since the forecasting of this wind takes into consideration the advection of cold air aloft over the Great Basin (Sergius 1952).

In some areas both foehn and bora winds seem to occur. For example, in a study of downslope gales at the east coast of Greenland, considered to be foehn-type winds, 5 out of 8 cases were found to be associated with a temperature rise, two with a temperature fall and in one case there was no appreciable change in temperature (Mirrlees 1934). Cadez (1953) in a discussion of the Kosava at Beograd, in terms of surface temperature changes, suggests that the wind, which is usually considered to be of the bora type, can at first be cool when cold air spills over the mountains to the east but can become quite warm later (over a period of several days) when the inversion on the windward side has been lowered to below the top of the mountains and advection and subsidence of warm air from above the inversion takes place. Kuettner (1940) studying the air flow during a Kosava (with strong surface winds) from a sailplane came to the conclusion that it has characteristics similar to those of the Foehn of the Sudeten Mountains in Germany (Kuettner 1939).

There is therefore no clear cut difference between the foehn and the bora. Downslope winds (including the occasional hurricane-force storms) may be warm or cold, from the descriptions in the literature it appears that both types can also occur in the same location, and a warming at lower levels may sometimes be associated with a cooling above. There are, furthermore, a number of characteristics which are common to both types of winds, such as the occasional violent and gusty storm and the relatively small areal extent of its effects, the lulls of the cold bora and the foehn pauses, and the local pressure fluctuations.

3. Air Flow Over Mountains - Observations and Theories

The nature of the air flow in the vicinity of mountains has received considerable attention for over 100 years, starting perhaps with the Austrian meterologist Hann and his now famous foehn theory proposed in 1866. Wild in Switzerland was probably the first (in 1867) to raise the question of the descent mechanism in the lee for the warm winds. Some of the theories are briefly mentioned in Section 2. In 1910 the Austrian von Ficker made his first daring balloon flight through the Foehn over Innsbruck.

With the increasing interest in gliding in the 1930's and the discovery of strong vertical motion in the lee of mountains, lee waves began to be observed and probed to an ever increasing extent in many parts of the world. A component of wind across the ridge and/or marked stability of the airstream at mountain top levels, with air of lesser stability above, have been found to be important factors in the development of lee waves in the lee of the

Riesengebirge by Kuettner (1939), in the central Swedish mountains by Larsson (1954), in the mountain country of Czechoslovakia by Foerchtgott (1957), in the Sierra Nevada by Holmboe and Klieforth (1957) and in the French Basses-Alpes by Gerbier and Berenger (1961), etc. In these observational studies the focus has generally been on the wave motion of the atmosphere and the associated cloud formation while surface winds are discussed only in passing, if at all. A reason for this is that many of the spectacular and beautiful clouds are a manifestation of small amplitude lee waves but are usually not present during strong downslope winds.

Air flow patterns across and east of the Colorado Front Range have been measured over the past several winters using aircraft as well as radar-tracked balloons. Results of these efforts have been published by Kuettner and Lilly (1968), Lilly and Toutenhoofd (1969), Vergeiner and Lilly (1970), Lilly (1971) and Lilly and Zipser (1972). In general, the results show two different types of air flow across the mountains. One is the regular lee wave type of flow. An example of this is seen in Fig. 3.1. The significant feature is the regular lee wave pattern consisting of periodic oscillations with relatively short amplitudes and wave lengths. Surface winds at the time of the cross section averaged about 22 m s^{-1} at the Divide (3749 m), 5 m s^{-1} at the upper slopes (3048 m) and 9 m s^{-1} in Boulder (1607 m) and the situation was not considered to be a wind storm in terms of the criteria used in the present study. The second observed pattern, shown in Fig. 3.2 (representing the potential temperature pattern during one of the wind storms considered in this study), has a single prominent

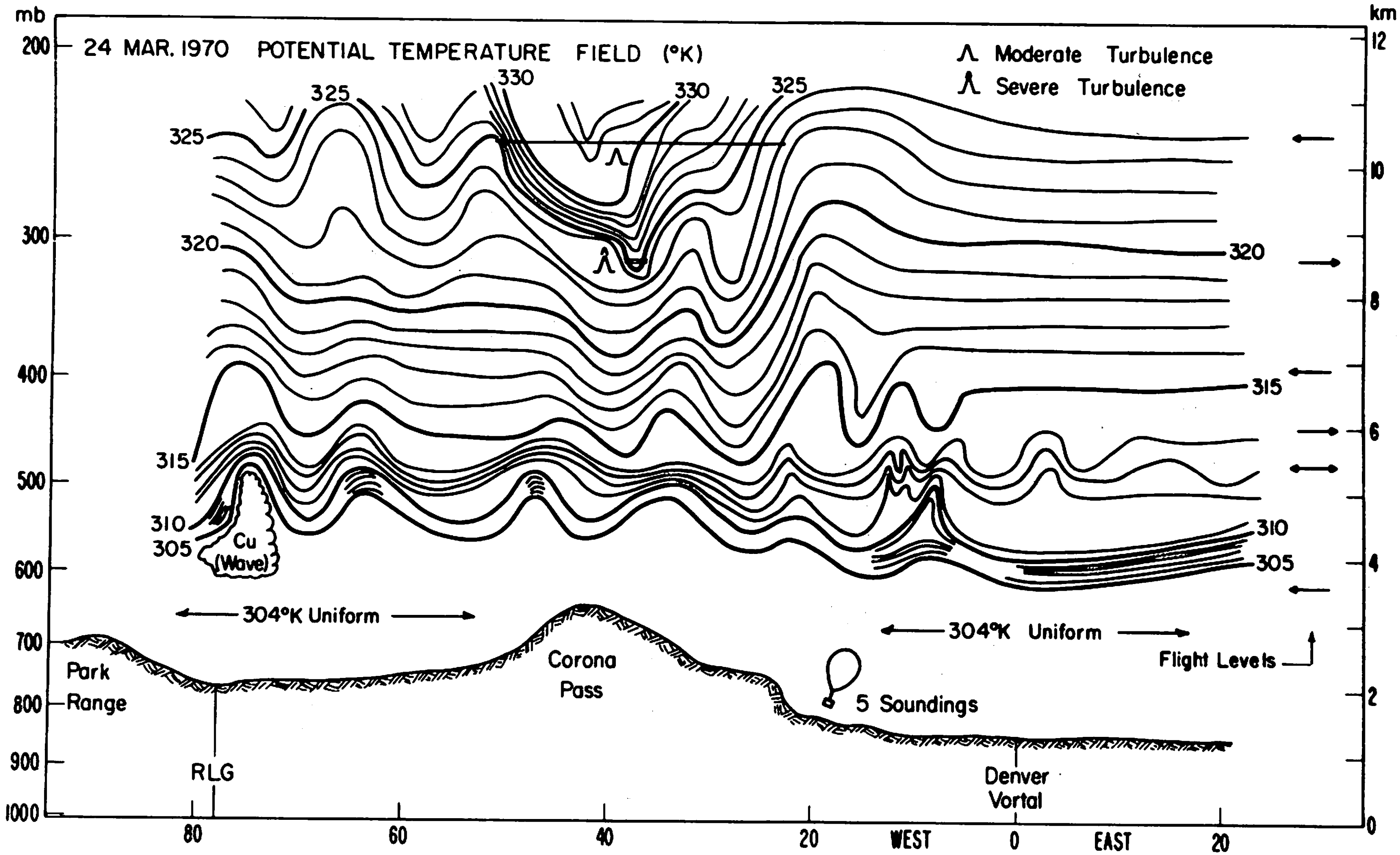


Fig. 3.1. Potential temperature cross section for 24 March 1970 over the Colorado Front Range between Kremmling and Denver, horizontal distance in nautical miles (presented at the Conference on Atmospheric Waves, 12-15 October 1971, Salt Lake City, Utah, by Zipser and Julian).

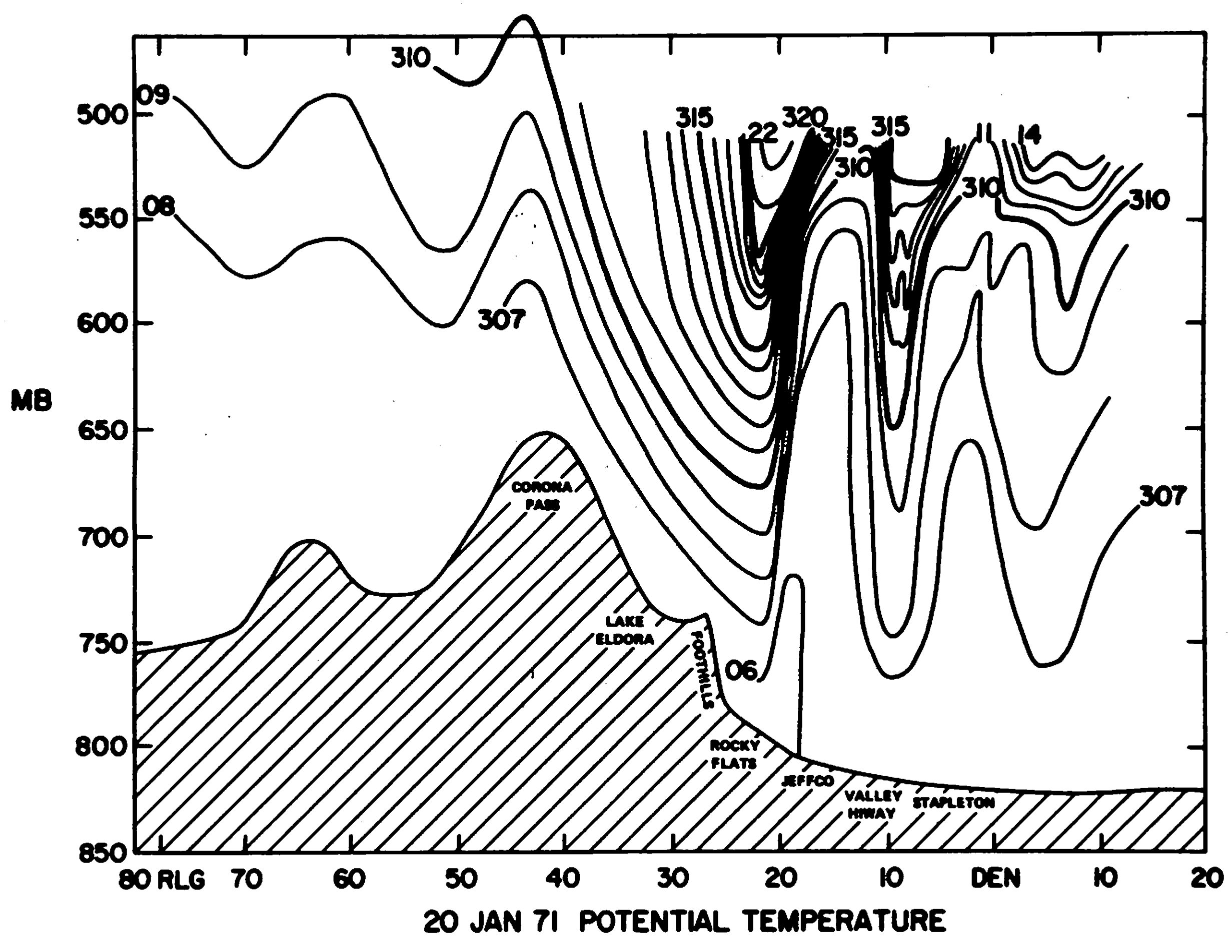


Fig. 3.2. Potential temperature cross section for 20 January 1971 over the Colorado Front Range between Kremmling and Denver, horizontal distance in nautical miles (presented at the Conference on Atmospheric Waves, 12-15 October 1971, Salt Lake City, Utah, by Zipser and Julian).

smooth downdraft regime just leeward of the Divide, an abrupt turbulent updraft to the east of it and some indication of lee waves further downstream. A similar pattern can be seen in Fig. 12.2 (Section 12) which represents conditions during one of the most severe wind storms in recent years but which occurred after the cut-off date for data collection in the present study.

The theoretical study of air flow over mountains began with Queney in 1936 (Alaka 1960) and has since been continuously developed by many workers including Lyra, Wurtele, Zirep, Scorer and others (comprehensive reviews of their work have been given by Corby (1954) and Alaka (1960)). Almost all theoretical approaches to the air flow over mountains have treated the disturbances as gravity waves: in a stably stratified air stream an air parcel subjected to a vertical displacement will attempt to regain its equilibrium position. In the so-called mountain wave or lee wave problem the source of the perturbation is the kinematic forcing by the mountain. Since the hydrostatic equations which describe the motion of any fluid are nonlinear, most of the theoretical investigations have made use of the perturbation method in which the motion is regarded as made up of an undisturbed basic flow on which is superimposed a small perturbation; this justifies neglecting the squares and products of the disturbance quantities and enables the equation to be linearized but restricts the results to small amplitude waves.

Scorer's (1949) contribution to lee wave theory was the introduction of a realistic air stream by providing for variations of lapse rate and wind shear with height. The combined effect of the vertical variations of the two variables on the dynamic properties of the air stream is given by vertical changes in the Scorer parameter

$$\ell^2 = \frac{g\beta}{U^2} - \frac{1}{U}\frac{\partial^2 U}{\partial z^2}$$

where g = gravitational acceleration
β = static stability = $\frac{1}{\theta}\frac{\partial\theta}{\partial z}$
θ = potential temperature
U = wind speed
z = height.

Scorer showed that for realistic lee waves to occur, an air stream must consist of at least two layers with stable air either forming the lowest layer or sandwiched between two layers of lesser stability. Consequently, ℓ^2 must be sufficiently greater in some lower layer than in the layer above and the difference must be greater the shallower the lower layer.

Mountain waves have long been suggested to be responsible for the generation of foehn-type winds. Variations in the ℓ^2 profile have been used in a number of cases as an indicator of such winds but the results have not always been very satisfactory (e.g., Lamb 1970).

To compute flow patterns in the lee of the Colorado Front Range, linear lee wave models have been developed by Danielsen and Bleck (1970) and Vergeiner (1971). Following the approach by Scorer (1949), the earth's rotation is neglected to simplify the equations; the assumption of laminar flow in lee wave theory represents another restriction. The computed patterns were found to be in close agreement with observed short waves; but, using upper air data for a number of Boulder wind storm cases, the results were not very satisfactory.

According to Yih (1965) and Scorer and Klieforth (1959) blocking upstream of a barrier will occur if the internal Froude number

$$F = \frac{v}{\sqrt{\frac{\Delta\theta}{\theta} gh}}$$

where v = velocity
g = gravitational acceleration
h = height of the mountains
θ = potential temperature

which is the ratio of the inertial force to the force of gravity for a given flow, is less than $1/\pi$ which corresponds to $h > \pi/\ell$. Scorer and Klieforth (1959) suggested that in that case the upper layer will descend to the surface on the lee side and speculated that this would lead to a warm foehn.

The π/ℓ criterion was tested by Lockwood (1962) in a study of foehn winds in the British Isles and in 4 out of 5 cases it was found that π/ℓ was slightly larger than the mountain height. Similarly, Beran (1967) used it in an analysis of Chinook winds in Colorado (defined in terms of dew-point spread, not wind speed) and found it to be a necessary but not a sufficient criterion for predicting leeward subsidence.

Apart from problematic theoretical assumptions, the reasons for some of the unsatisfactory results from the application of either ℓ^2 or π/ℓ are the difficulties associated with calculating the Scorer parameter from the available upper air data. The wind shear term, for instance, is usually neglected although it is not always negligible. Another problem lies in the definition of the foehn which is usually thought of in terms of sudden temperature rises at the surface. A foehn thus defined may be the result of a number of mechanisms (see, e.g., Glenn 1961) and may or may not be accompanied by strong surface winds.

The 'hydraulic jump' model provides another approach to the problem of air flow across mountains which does not assume laminar flow but actually predicts turbulent motion in certain regions, and is not restricted by assumptions of short amplitudes and wavelengths. The phenomenon, essentially a shock wave, can be observed in a channel at the foot of a weir where water arrives with a high velocity. Long familiar to engineers, it was brought to the attention of meteorologists by McGurrin in 1942 when he described a stationary jump in the height of a fog bank (Freeman 1951). Freeman (1948) applied the theory to waves in the equatorial easterlies, Tepper (1950) to squall lines, and Fujita (1955) showed thunderstorm induced pressure jump lines; one of the most outstanding examples is perhaps the jump in surface pressure of 6.1 mb within 2 minutes described by Williams (1948) in association with a squall line. In 1953 Schweitzer suggested that the hydraulic jump model might be applicable to strong Foehn winds in the European Alps.

According to hydrodynamic theory (Schweitzer 1953; Long 1954; Ball 1956; Kuettner 1959) flow across a barrier (Fig. 3.3) may be either subcritical (tranquil flow), critical, or supercritical (shooting flow). The critical velocity is a function of the height of the flow, i.e.,

$$v_c^2 = g\,h$$

where v_c = critical velocity
h = height of flow above mountain top level
g = gravitational acceleration.

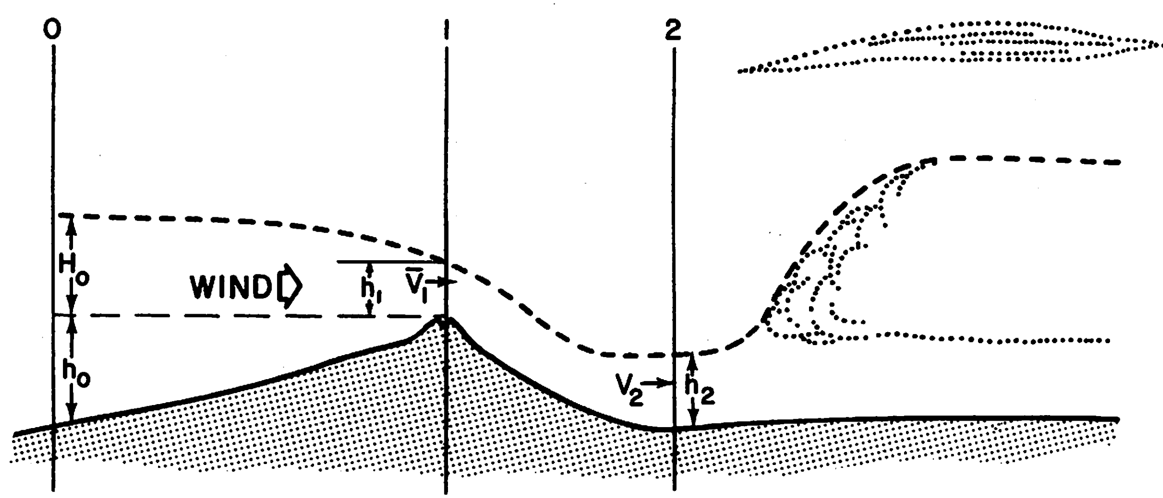


Fig. 3.3. Schematic diagram of hydraulic air flow over a mountain range (after Kuettner 1959).

Along the lee slopes, critical flow could be accelerated to a supercritical velocity. Since the critical velocity is also the maximum propagation speed of long surface waves, such waves can travel upstream only in the subcritical but not in the supercritical area. The wave front will therefore steepen until it breaks down at the critical point. In this way the turbulent hydraulic jump is formed at the transition from shooting to tranquil flow. The mean flow kinetic energy which is lost at the jump appears either as turbulent kinetic energy and/or is radiated away by a stationary wave system situated downstream of the jump (Binnie and Orkney 1955).

Applying the hydraulic jump theory to the atmosphere, the free surface of water flowing in an open channel changes to an interface (inversion) between a colder (denser) air mass below and a warmer (lighter) one above. This reduces the gravitational force which must be replaced by a 'modified gravity' given by

$$\gamma = \frac{\Delta\theta}{\theta} g$$

where θ = potential temperature
g = gravitational acceleration.

This modified analogy still contains some unrealistic features such as the discontinuity in the dry-adiabatic lapse rate at the boundary between the layers (the inversion is thus a layer of zero thickness) and the neglect of the continuous stratification of the flow above the inversion.

In 1953 Schweitzer pointed out the similarity between the hydraulic jump model and the strong Foehn winds of the European Alps. Ball (1956) applied the theory to strong cold winds blowing down the slopes of Antarctica's large ice cap. Arakawa (1969) used hydraulic jump theory to explain a foehn-type wind on Hokkaido, Japan. Kuettner (1959) suggested the hydraulic jump as a possible mechanism for both bora and foehn-type winds. He also introduced the heated jump according to which additional energy in the form of sensible heat, imparted to the air as it descends in the lee, would permit the height of the jump to exceed that of the crest cloud.

The atmospheric phenomenon analogous to the hydraulic jump is usually referred to as a pressure jump since its passage over a station causes a sudden change in depth of the heavier lower layer and a corresponding sudden change in surface pressure. Thus, for downslope winds, when the jump moves downstream past the station, the onset of strong surface winds is accompanied by a pressure drop while an upstream moving jump would result in high winds followed by a pressure rise. Temporary changes in atmospheric conditions upstream may cause the jump to move up or down the slopes resulting in temporary lulls or pauses at a station.

Houghton and Isaacson (1968) extended the hydraulic jump theory to a three-layer model, each layer having a constant potential temperature (thus being separated from the next by an inversion of zero thickness) and the two lowest layers having a constant wind speed while the top layer has no mean wind velocity (a very crude representation of atmospheric conditions). The flow pattern for a day for which upper air data from aircrafts and balloons existed for the eastern Colorado area was computed. The results, superimposed on the observed cross section, are shown in Fig. 3.4. Since the interface (inversion) appeared to be between about 4880 and 5180 m, they performed two integrations of the system. The results show that a lowering of the inversion causes the jump to

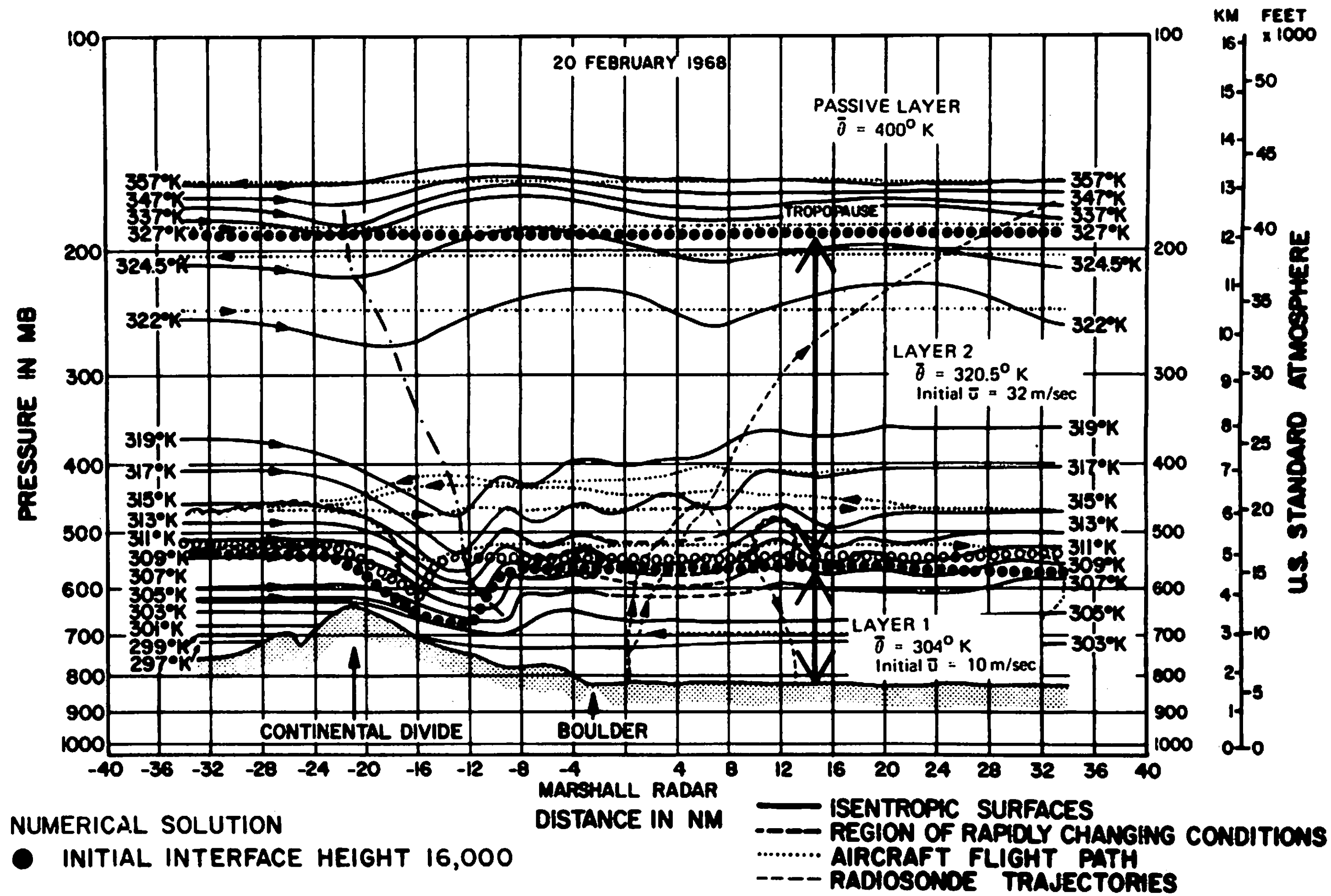


Fig. 3.4. Superposition of two steady state solutions on isopleths of observed potential temperature (from Houghton and Isaacson 1968).

move downstream from close to the summit to halfway down the slopes. Hourly wind speeds in Boulder at the time of the cross section were less than 2 m s^{-1} while those at the upper mountain stations reached 12 m s^{-1} which agrees with the observed and theoretical air flow in Fig. 3.4.

Klemp and Lilly (1973) have very recently re-examined the lee wave problem and produced a solution involving a three-layer atmosphere. They have shown that because of the interaction between the three layers the maximum surface velocity, for long waves, is given by

$$U_m = \frac{(N_o U_o)(N_2 U_2)}{(N_1 U_1)^2} \quad (N_o h)$$

where θ = potential temperature
g = gravitational acceleration
h = height of the mountain
U = mean wind speed
z = vertical height
$N = \sqrt{\frac{g}{\theta}\frac{\partial \theta}{\partial z}}$

subscripts o, 1, 2 refer to the bottom, middle and top layer respectively.

Maximum amplification, i.e., optimum tuning of the three layers, occurs with a semi-infinite top layer and the two lower layers each one-quarter vertical wavelength deep, but with large wavelength differences between them. This would be achieved with a low stability in the middle layer (middle troposphere), high stability in the bottom and top layer (i.e., an inversion or isothermal layer at low levels and a stable stratosphere). For wind speeds in the lowest isothermal layer equal to 20 m s^{-1}, its depth should be about 1.5 km thick to produce optimum conditions (which is somewhat too thick to be realistic). For high winds at the surface, a high mean speed at mountain top levels (lowest layer) and velocities not too high in the middle layer are desirable. Numerical results have suggested that deviations from optimum conditions could result in changes in the location of the surface wind maximum on the lee slopes. Results from the non-linear model have been found to confirm those obtained through linear theory. Work on this basic model, presently in progress, permits changes in the stratification of the lowest layer from one of constant stability to a more realistic one having a shallow adiabatic layer close to mountain top and a sharp inversion above. The sharper inversion case tends to produce stronger maximum surface winds.

In summary, most of the early work on air flow over mountains concentrated on small disturbances not applicable to strong downslope winds and attempts to use the approach have furthermore been hampered by the problem of definitions of the foehn. The hydraulic jump model, in spite of several unrealistic features, appears to explain a number of the observed characteristics of downslope winds. The Klemp and Lilly model, in its modified form which in the lowest layer encompases the whole spectrum of atmospheric stratification, is a new attempt to realistically approach the downslope wind problem. However, observational analyses have so far been insufficient to determine the exact nature of the air flow across and in the lee of the Divide during wind storms in Boulder. For this reason, the term 'lee flow disturbance' rather than trough or jump will be used in this study.

For the purpose of the present study the most important conclusions are that there is no doubt about the significance of the upstream air flow characteristics, particularly the existence of a stable layer above mountain top levels, and that theory suggests the movement of deep lee troughs or hydraulic jumps in response to changes in upstream air flow characteristics. Such movements would explain pauses or lulls in surface wind and large fluctuations in surface pressure. All of these surface and upper air features have been observed with many warm and cold downslope winds around the world and are also characteristic of Boulder's wind storms, as the results presented in the following chapter show.

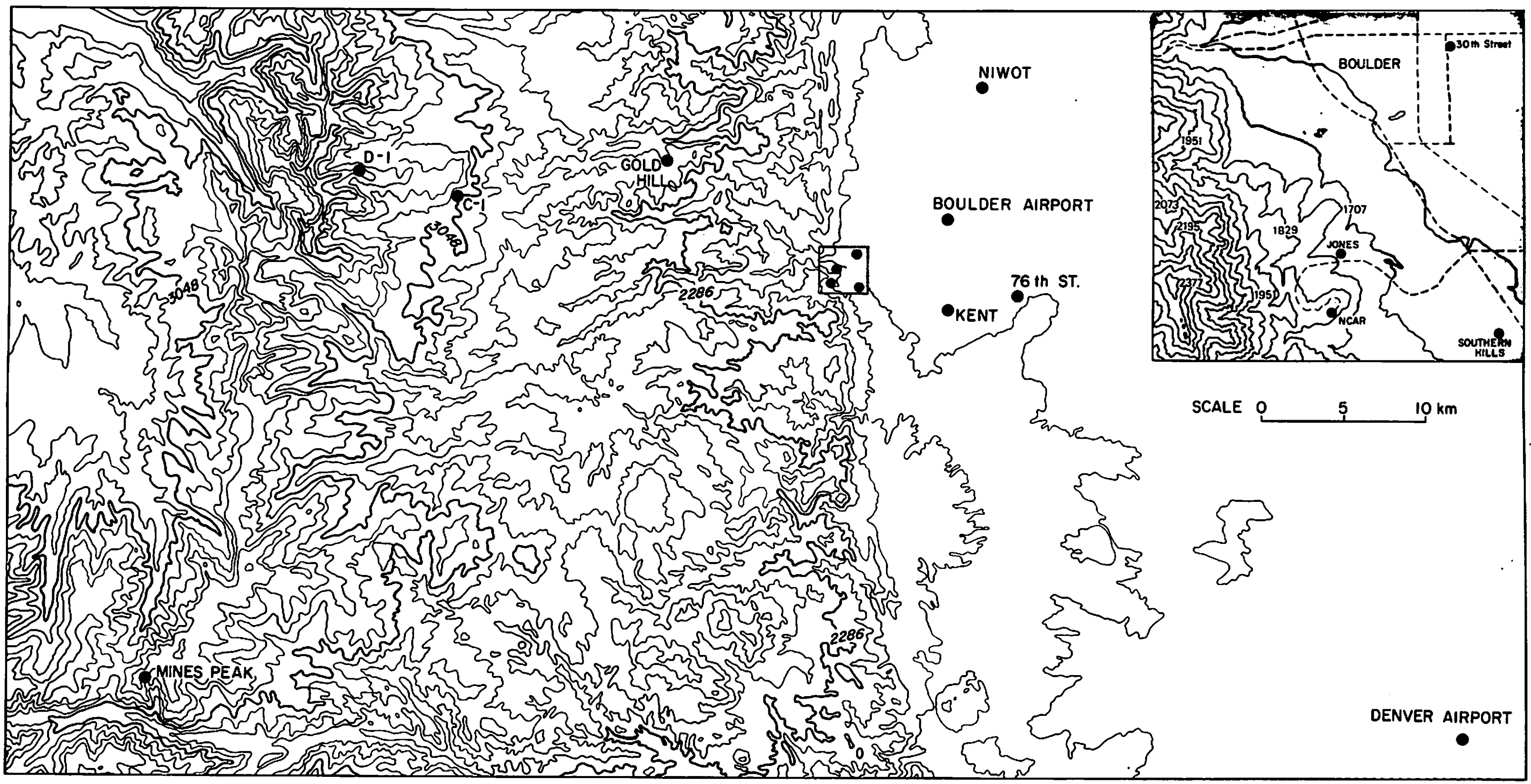
CHAPTER II

DATA AND METHODS

4. Data

a. Basic Data. The surface temperature and humidity data used in this study consist of records from a network of weather stations operated by the University of Colorado's Institute of Arctic and Alpine Research (INSTAAR), the National Center for Atmospheric Research (NCAR), the National Weather Service Forecast Office at Denver (NWSFO), and some stations operated by the author especially for this study since the scarcity of continuous temperature data, and particularly the lack of humidity data, around Boulder was considered serious. The location of these stations is shown in Fig. 4.1 and detailed descriptions are given in Table 4.1.

It must be noted here that while thermographs are quite accurate, hygrographs are seldom as precise as is to be desired, particularly when the humidity is low. Furthermore, both types of instruments give a good deal of trouble at low temperatures, such as stopping of clocks, snow blowing into the shelter and clinging to the instruments, especially when clinging to the hair element. Another problem is caused by the slow response of the hygrograph. It takes about 5 minutes for the instrument to indicate 90 per cent of a sudden change at 25°C and up to 10 times longer at -10°C (Wexler and Brombacher 1951). Spot checks of relative humidity readings made with a ventilated psychrometer are also somewhat unreliable at low temperatures due to difficulties in determining the correct wet-bulb temperature, small fractions of a degree error can cause an indication of twice the actual humidity (Ruskin 1963). Temperature, and particularly humidity readings,



NIWOT
GOLD HILL
D-1
C-1
3048
2286
BOULDER AIRPORT
76 th ST.
KENT
MINES PEAK
DENVER AIRPORT
SCALE 0 5 10 km
BOULDER
30th Street
1951
2073
2195
2377
1829
1707
JONES
NCAR
SOUTHERN HILLS

Table 4.1. Description of weather stations (temperature and humidity)

Station	Operating agency	Elevation (m)	Exposure (*)	Instrumentation
D-1	INSTAAR	3,743	ridgetop - open tundra	Bourdon tube and hair hygrothermograph (BENDIX, USWB Spec. No. 450.8202)
C-1	INSTAAR	3,048	ridgetop - forest clearing	
B-1	INSTAAR	2,591	ridgetop - forest clearing	
A-1	INSTAAR	2,195	ridgetop - forest clearing	
NCAR	NCAR	1,838	at foot of Flatirons on 6.1 m mast above 16.8 m building	Diode-junction type thermometer (NCAR design)
PSRB#1 (30th St.)	INSTAAR(**)	1,608	on roof of 10.7 m building	Bourdon tube and hair hygrothermograph (BENDIX, USWB Spec. No. 450.8202)
Boulder Airport	NCAR(**)	1,604	E of runway and 150 m N of airport building	Bourdon tube and hair hygrothermograph (BELFORT, USWB Spec. No. 450.8202)
Niwot	NCAR(**)	1,564	on grazing land about 100 m N of farm building	Bourdon tube and hair hygrothermograph (BENDIX, USWB Spec. No. 450.8202)
Denver Airport	NWSFO	1,610	E of runway - open terrain	H061 hygrothermometer (liquid-filled type element)

(*) instruments, except NCAR, are housed in a standard shelter, 1.2 m above ground
(**) stations operated by author during the winters 1969/1970 and 1970/1971; instruments supplied by the respective agencies.

are therefore less reliable during the cold season and at high elevations; and errors in the humidity readings may easily reach ± 10 per cent.

During the abstraction of the temperature and humidity data for this study (only Denver's data were available in tabulated form) the thermograms and hygrograms were therefore corrected not only for time but also for some of the more obvious temperature and humidity errors by comparing chart readings with notes taken at times of setting and removal of charts, with precipitation records, observation notes, by checking the 100 per cent readings on the hygrograms and by evaluating the general appearance of the humidity trace (snow in the hair element, for instance, causes the trace to be a relatively straight line gradually dropping below the 100 per cent mark). After discarding 'unsalvageable' data, hourly averages, centered on the hour, were determined to the nearest degree and per cent. Degrees Fahrenheit were converted to degrees Celsius, again to the nearest degree in view of the probable accuracy of the data.

Temperature observations on top of buildings are under normal conditions of doubtful value because of rapid variations in the vertical and the effect of the building itself. However, adiabatic conditions can be assumed for periods with strong winds. With a dry-adiabatic lapse rate the difference in temperature between the ground and the instrument on the roof of PSRB#1 is 0.1°C and 0.2°C for the NCAR instrument. For the purpose of this study, such differences are insignificant.

The main body of surface wind data was provided by NCAR which has set up a network of 7 All-Purpose Recording Systems in the Front Range-Boulder Valley area since 1965. Before 1968 the location of a number of the instruments was changed every winter to investigate local variations in the wind storms. The location of the stations used in the present study are shown in Fig. 4.1 and detailed descriptions are given in Table 4.2.

As is the case with hygrothermographs, there are certain problems inherent in anemometers. The distance constant (the length of an air column that must pass after the release of the cups for them to reach 63 per cent of the terminal velocity) is relatively high for the All-Purpose Wind Recording Systems as compared to the MRI instruments, for instance (see Table 4.2). A high distance constant results in a large degree of smoothing. For instance, for a mean speed of 15 m s^{-1} sinosoidal fluctuations with a period of 7 seconds are resolved with an accuracy of 90 per cent while fluctuations of the order of two seconds are resolved with an accuracy of less than 50 per cent (Gill 1967). Also due to the high inertia, the anemometer will register too high a mean speed in a gusty wind because the cup wheel accelerates more than it decelerates; for ordinary instruments this can reach 10 per cent (Middleton and Spilhaus 1953). It must be pointed out, however, that the method of estimating mean speeds by eye from a chart record is itself probably subject to uncertainties of the same order.

During the abstraction of the wind data the timing of the wind charts was checked against operator's notes (unfortunately, these

Table 4.2. Description of anemometer stations

Station	Operating agency	Elevation (m)	Exposure (*)	Instrumentation
Mines Peak	USFS	3,808	exposed summit of Mines Peak on 11.9 m mast	DC generator type 3-cup anemometer system
D-1	INSTAAR(**)	3,743	ridgetop - open tundra on 6.6 m mast	DC generator type 3-cup anemometer system, continuous speeds up to 150 mph
C-1	INSTAAR(**)	3,048	ridgetop - forest clearing on 7.3 m mast	All-Purpose Wind Recording System (SCIENCE ASSOCIATES, INC.) continuous speeds up to 100 mph instantaneous directions at one-minute intervals and to 8 compass points starting speed about 1 m s^{-1} distance constant about 7.9 m
Gold Hill	INSTAAR(**)	2,591	ridgetop - forest clearing surrounded by young aspen	
Jones	NCAR	1,768	below Flatirons, on the NE slopes of Table Mesa	
Southern Hills	NCAR	1,676	on roof of 7.6 m building	
30th Street	NCAR	1,608	on roof of 10.7 m building	
Kent	NCAR	1,622	about 15 m E of a one-story family house	
76th Street	NCAR	1,641	about 15 m E of a one-story family house	
Boulder Airport	NCAR(***)	1,604	E of runway and about 150 m N of airport building, open terrain, on 2.0 m mast	MRI system records run of wind and continuous wind direction distance constant about 2.4 m
Niwot	NCAR(***)	1,564	grazing lands - about 100 m N of farm building, on 2.0 m mast	
Denver Airport	NWSFO	1,610	E of runway - open terrain on a 9.1 m mast	F420C system, DC generator type 3-cup anemometer, starting speed about 1 m s^{-1}

(*) instruments are installed on top of a 3.4 m support, unless otherwise noted
(**) wind instruments on loan from NCAR
(***) stations operated by author during the winters 1969/1970 and 1970/1971; instruments supplied by the respective agencies

were often missing). Whenever the timing was out by more than three minutes in the same sense at both ends of the chart, an adjustment was made in 5-minute steps since it is the length of the averaging period used in this study. Wind speed was corrected whenever an obvious and persistent change in the zero reading was found. Problems also arose because the 100 mph chart scale was obviously exceeded in some cases but had to be recorded as 100 mph. Occasional malfunctioning of the direction indicator system was the cause of several problems such as more than 5 recordings per 5-minute interval, two adjacent direction arms stuck together and thus recording simultaneously, no recording from one or more arms; these were corrected as objectively as possible but in some cases the direction data for that period and station had to be completely rejected.

Five-minute mean (visually estimated), maximum and minimum speeds and the 5 directions per 5-minute interval were abstracted using NCAR's Datagrid graphic digitizing system (BENDIX). This electronic system records specific points on the chart on magnetic tape in terms of x- and y-distances from a given reference point. The raw data were first run through a prepass program (provided by NCAR) to rotate the x and y axes and remove error messages. A second program (especially written for this study because of the variability of chart dimensions) converted the x and y distances into wind speed (in meters per second) and direction (in degrees), averaged the 5 direction values for the 5-minute intervals and punched the results onto cards.

To supplement the network of 7 stations, additional wind data for a number of other stations were obtained from the U.S. Forest Service (USFS), the National Weather Service Forecast Office in Denver (NWSFO) and abstracted from charts from stations operated by the author. (The locations of these stations are shown in Fig. 4.1 and detailed descriptions are given in Table 4.2.) Only hourly wind data from these stations were used for a 'meso-scale' analysis of the wind storm periods, as defined in Section 6, since hourly data are less affected by differences in instrumentation. For the network of 7 All-Purpose anemometer stations hourly mean speed and direction (centered on the hour) were calculated from the 5-minute mean speed and direction values. Since the period covered by the hourly data was usually longer than the period covered by the 5-minute data, additional hourly mean speeds were estimated by eye and mean directions were calculated from digitized data (since direction is more difficult to estimate from these charts).

Differences in instrumentation and anemometer heights are an important problem in obtaining comparable results, but differences in exposure may in some cases be of greater importance and are practically impossible to correct for. Anemometers in forested areas, on slopes, near houses, on top of asymmetrical buildings, for instance, represent such a variety of different and non-standard exposures that a reduction of the data to a standard height could be meaningless. However, the existence of a relatively dense network of stations in an area of strong downslope winds is relatively unique (at least on this continent) and thus valuable.

A determined effort was therefore made to utilize the available data.

For the reduction of wind data from all stations with anemometer heights different from 3.4 m (except for Southern Hills and 30th Street), the simple power law was used

$$\frac{v_1}{v_2} = \left(\frac{z_1}{z_2}\right)^p$$

where v_1 = wind speed at the upper level, z_1
v_2 = wind speed at the lower level, z_2.

The exponent, p, depends upon the vertical temperature gradient and surface roughness. A value close to 1/7 is considered to be appropriate for smooth open country and an adiabatic lapse rate, which can be assumed to exist under conditions of high winds (Sutton 1953; Blackadar 1960).

For the Southern Hill High School and 30th Street locations the effect of the buildings on the anemometers on the roof was unknown. To obtain empirical formulae, two identical contact cup anemometers (BELFORT, U.S.W. Spec. No. 450.6104, starting speeds of about 1 m s^{-1}, giving a contact for each mile of wind) were used. One was installed next to the anemometer on the roof, the other on a 3.4 m support on the ground next to the building. The two travelling anemometers were connected so that they recorded simultaneously on the same chart to facilitate comparisons.

In both cases, the Southern Hills and the 30th Street site, the ground anemometer was, unavoidably within 20-50 m of some buildings to the north and south, but without obstructions to the west and east. At the Southern Hills site the test was run from 8 December 1970 to 11 January 1971, the instruments were then moved to the 30th Street site where the test was run from 11 January 1971 to 1 March 1971.

Due to the limitations imposed by adjacent buildings, periods with westerly winds (WNW, W, WSW) only were used for the comparison. Furthermore, periods with wind speeds of less than 2.2 m s^{-1} at the ground were discarded because of uncertainty in wind direction at low speeds. Both these restrictions were considered acceptable since the results were intended to be applied to strong westerly winds. The actual comparison was done by counting the number of contacts (and estimated fractions) made by the two anemometers over identical 30-minute intervals. The length of this period was arbitrarily chosen considering the problem caused by the possibility of large changes in mean speed over periods of less than one hour and estimations of runs of wind for short periods of time. Due to the direction and speed restrictions the tests at the two sites provided a total of 98 half-hours of useful data each.

The results, shown in Fig. 4.2, suggest a linear relationship. Simple regression and correlation analysis was performed on the two sets of pairs which gave the following results

for Southern Hills High School:

$$y = -0.32 + 0.91x \qquad r = 0.98$$

for 30th Street:

$$y = -0.52 + 0.92x \qquad r = 0.99$$

where y = ground speed
x = roof speed

applicable over the range of 3 to 22 m s^{-1} mean hourly wind speed. These equations are not applicable to peak wind speeds. Air in large gusts tends to retain its velocity; maximum gusts therefore decrease downward less rapidly than mean wind speeds (Deacon 1955).

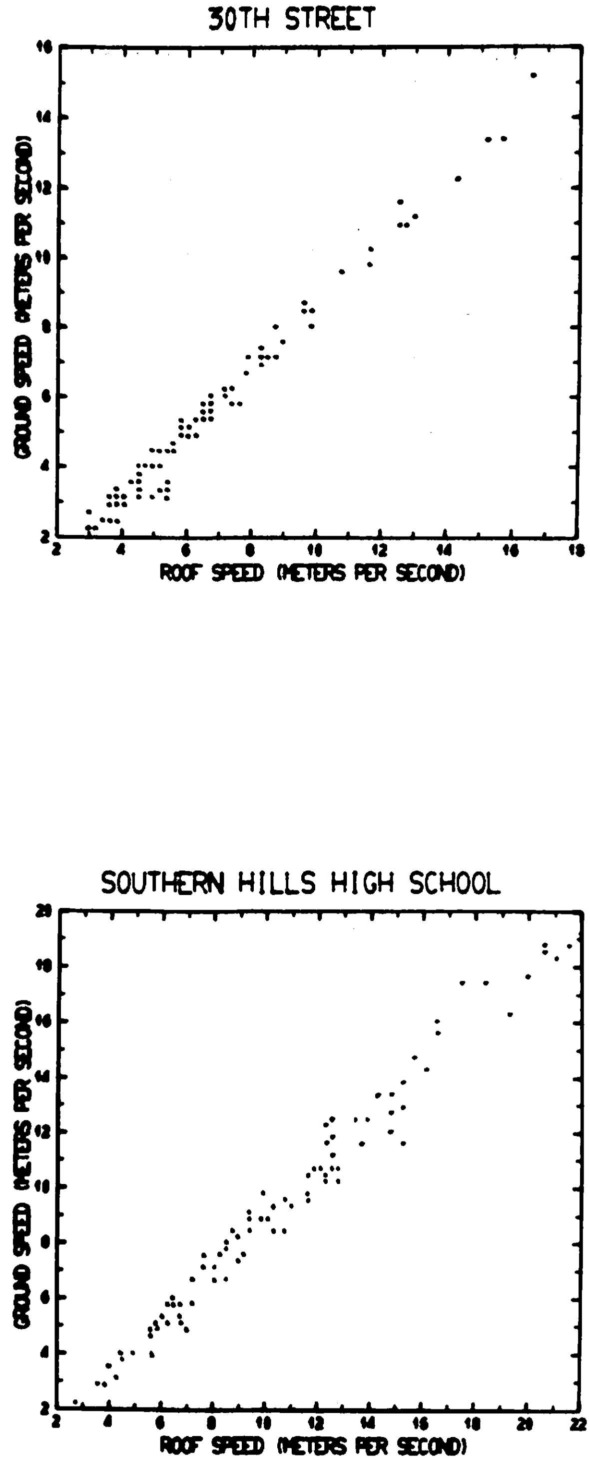


Fig. 4.2. Ground speed as a function of roof speed.

Although the data sets do not represent true time series, but a collection of disconnected sections, a significant amount of autocorrelation was found to exist in the data (r_1 varied between 0.66 and 0.57 which is significant at the 0.01 level). To estimate the effective number of independent observations to be used in testing the significance of the correlation coefficients, a formula due to Bartlett (Quenouille 1952) was used:

$$N/(1 + 2r_1r_1' + 2r_2r_2' + \ldots)$$

where N = actual number of observations
$r_1, r_2, \ldots$ = autocorrelation coefficients for series 1
$r_1', r_2', \ldots$ = autocorrelation coefficients for series 2.

From the above equation the estimated number of independent observations were 33 for Southern Hills and 45 for 30th Street. These give significance at the 0.01 level to the above correlation coefficients. The two equations were therefore accepted as excellent approximations in reducing the mean roof wind speeds to the 3.4 m level.

A summary of the reduction factors for all stations with anemometer heights different from 3.4 m is given in Table 4.3.

Regarding surface pressure data, only one set of homogeneous long-period pressure records was found for the Boulder area. The barograph has been operated by NCAR at the Table Mesa Laboratory since 1965. Since the instrument had originally been set to read Denver's altimeter setting, it records only relative changes, and absolute values would have to be determined by comparison with other instruments. Because of the occasional resetting of the barograph and possible drift over the years, absolute values were determined for every period in which wind cases occurred. Frequently checked barographs have been kept intermittently at various locations, such as 30th Street, Boulder, and Marshall, Colorado. Using such records, and selecting absolute highs and lows and periods of relatively constant pressure, an average difference between NCAR and the comparison station was obtained. A theoretical difference was calculated from the known elevation of the two locations and the U.S. Standard Atmosphere. The difference between the theoretical and calculated value gives the adjustment factor which was added to the NCAR pressure data. This varied between -209.0 mb and -202.5 mb over the period 1969/1971.

Table 4.3. Summary of anemometer height correction factors

Station name	Anemometer level	Correction factor
Mines Peak	11.9 m	0.83
D-1	6.6 m	0.91
C-1	7.3 m	0.89
Southern Hills	11.0 m	0.91
30th Street	14.0 m	0.92
Boulder Airport	2.0 m	1.08
Niwot	2.0 m	1.08
Denver	9.1 m	0.87

In addition, station pressure data were obtained for Denver from the National Weather Service Forecast Office at Denver and for Eagle (a station on the windward slopes of the Colorado Front Range, at an elevation of 1980 m, about 135 km west of Boulder) from the Colorado State Climatologist in Denver.

All pressure data are hourly values covering for each of the 20 wind storms the period defined as mesoscale in Section 6.

Hourly <u>precipitation</u> and <u>upper air</u> data were obtained from the routinely published records.

b. Derived Data. <u>Pressure values</u> for stations other than Eagle, NCAR and Denver were needed to calculate the potential temperature and were computed from the equation

$$P_2 = P_1 \, e^{\frac{-g}{R\bar{T}}(z_2 - z_1)}$$

where P_1 = pressure at lower station
P_2 = pressure at upper station
z_1 = elevation of lower station
z_2 = elevation of upper station
$\bar{T}$ = mean absolute temperature of the stations
R = gas constant.

Station pressure for the mountain stations was calculated by working upward from NCAR to A-1, from A-1 to B-1, etc. The calculations for the Boulder stations always started with NCAR because of frequent zero temperature differences between the Boulder stations.

<u>Potential temperature</u>, θ, was calculated using the relationship

$$\theta = T \left(\frac{1000}{P}\right)^{0.288}$$

where T = absolute temperature
P = station pressure.

The <u>mixing ratio</u>, w, which is the ratio of the mass of vapor present to the mass of dry air containing it, is given by

$$w = r\,w_s$$

and

$$w_s = \varepsilon \frac{e_s}{p - e_s}$$

where w_s = saturation mixing ratio
r = relative humidity
p = pressure
e_s = saturation vapor pressure
ε = ratio of molecular weights of water and dry air.

Since the hair element of the hygrograph measures the relative humidity with respect to liquid water at all temperatures (Middleton and Spilhaus 1953), e_s was calculated using the equation

$$\ln \frac{e_s}{6.105} = 25.22 \frac{(T - 273)}{T} - 5.31 \ln \frac{T}{273}$$

where T = absolute temperature
e_s = saturation vapor pressure

which is accurate in the range $-20°C \leq t \leq 50°C$ (Berry *et al.* 1945, p. 343).

c. Accuracy of Data. Within the range of temperatures and pressures encountered in this study, the maximum error in the calculated pressure resulting from a 2°C error in the mean temperature between stations is less than 0.5 mb, and the maximum error in the potential temperature resulting from a 5 mb error in the NCAR pressure data is less than 1.0°C. An error of 1.0°C in the station temperature will affect the calculated saturation mixing ratio by about 8 per cent which would amount to 0.49 gm kg^{-1} at 0°C and 650 mb pressure. A 10 per cent error in relative humidity would result in a difference in mixing ratio of 1.3 gm kg^{-1} at 15°C and 850 mb pressure.

A check of the actual mixing ratios at the mountain stations during downslope winds showed irregular and unreasonable differences. Two stations may, for instance, be found to have very similar ratios while a station between them would only record half that value. It was therefore concluded that the relative humidity data were too unreliable to give acceptable mixing ratio data.

5. Methods

Since the size of the sample was relatively small (20 storm cases) and the uncertainty of some of the data was known (especially that of the twice-daily upper air soundings), the use of sophisticated statistical techniques was not considered to be appropriate. Scatter diagrams, median values and t-test were found most useful.

To test the difference of means for small samples, the confidence limits were calculated using

$$t = \frac{x_1 - x_2 - (u_1 - u_2)}{s_{12}} \quad \text{where} \quad s_{12} = \frac{N_1 s_1^{\,2} + N_2 s_2^{\,2}}{n_1 + n_2} \left(\frac{N_1 + N_2}{N_1 N_2}\right)$$

and to test whether a mean is significantly different from zero, the confidence limits were calculated using

$$t = \frac{(x - u)\, n^{1/2}}{s} \qquad \text{where} \qquad u = 0.$$

To test the difference of variances, the ratio

$$F = \frac{N_1 s_1^{\,2}}{n_1} \bigg/ \frac{N_2 s_2^{\,2}}{n_2}$$

was used. For further details see Kenney and Keeping (1954).

Regarding the base of the stable layer, discussed in Section 13, to reduce the effect of a few extreme values, the center of the modal class instead of the mean was used and differences between groups of observations were tested using the Mann-Whitney rank test (for details see Snedecor and Cochran 1967).

For the comparison of regression lines the analysis of covariance was used to test for homogeneity of variance, slope and elevation. Regression lines are, for instance, different if they have the same slope (i.e., are parallel) but differ significantly in elevation. For details see Snedecor and Cochran (1967, pp. 432-436).

In general, 'significant' differences in results in Chapter III imply significance at the 5 per cent level (also indicated by the symbol (*)); except in a few important cases the 5 per cent level was the only criterion used since this greatly simplified the computer calculations.

The processing of the data was done on the University of Colorado's CDC 6400 and NCAR's CDC 6600 computers.

6. The Wind Storm Cases

For the purpose of this investigation a wind storm period was defined as one during which maximum speeds at the Boulder stations (excluding 76th Street, the one furthest to the east) exceeded 22 m s^{-1}, with at least one station recording a gust of over 32 m s^{-1}, a hurricane-force wind. Because of differences in anemometer heights, this threshold value was increased to 34 and 35 m s^{-1} for the 30th Street and Southern Hills locations, although these may be overestimates because of the less rapid downward decrease of gust speed than mean speed.

Because of differences in the length of the records for the anemometer stations, the study was limited to the three winters 1968/1969, 1969/1970 and 1970/1971. Only the winter months were considered because some of the NCAR anemometers had been installed sometime in the early winter (about November) and taken down again in the late spring (about May) to be used for other research projects during the summer. The length of the 'winter period' is therefore variable and a wind storm early in the fall may not be included. However, these are infrequent according to Julian and Julian (1969) who, from a search of newspaper and other records as far back as November 1906, found November to March to be the most important months for strong winds in Boulder, with the largest number occurring in January (see Section 7). During the summer period strong wind storms do occur, although very infrequently, and they may also be of a different nature. Whiteman and Whiteman (1973, pers. comm.) in their newspaper survey found storm occurrences in all months except July.

The length of the storm periods for the small-scale analysis (using 5-minute wind data) was determined such that the speed at the beginning and end approached zero at the Boulder stations. For the mesoscale investigation (using hourly data), additional hours at the beginning and end were included to obtain information regarding conditions at the mountain stations, i.e., whether or not strong winds were blowing at those stations for some time before and/or after their occurrence in Boulder. In cases where the wind set in, or terminated, almost simultaneously everywhere, few if any hours were added.

Using the above criteria, a total of 20 wind storms occurred over the three winters. Their dates and the periods covered by the 5-minute and hourly data are shown in Table 6.1. Table 6.2 lists the maximum gusts recorded at the 7 stations during the 20 wind storms.

Table 6.1. Dates of the 20 wind storm cases
(Top : period covered by 5-minute data)
(Bottom : period covered by hourly data)

1.	7 January	12:00	-	8 January	12:00	1969
	6	1:00	-	8	12:00	
2.	29 January	16:00	-	30 January	16:00	
	29	13:00	-	30	19:00	
3.	31 January	3:00	-	31 January	12:00	
	30	21:00	-	31	15:00	
4.	18 March	21:00	-	19 March	18:00	
	18	21:00	-	19	18:00	
5.	7 April	0:00	-	8 April	16:00	
	6	21:00	-	8	19:00	
6.	25 January	0:00	-	26 January	10:00	1970
	24	18:00	-	26	18:00	
7.	3 February	3:00	-	4 February	5:00	
	2	17:00	-	4	8:00	
8.	16 February	23:00	-	17 February	8:00	
	16	15:00	-	17	10:00	
9.	21 November	2:00	-	21 November	18:00	
	20	10:00	-	21	24:00	
10.	24 November	22:00	-	25 November	18:00	
	24	16:00	-	25	21:00	
11.	28 November	10:00	-	29 November	10:00	
	28	6:00	-	29	15:00	
12.	30 November	7:00	-	1 December	19:00	
	30	4:00	-	1	22:00	
13.	31 December	8:00	-	1 January	4:00	1970/
	31	1:00	-	1	8:00	1971
14.	20 January	12:00	-	20 January	24:00	1971
	20	6:00	-	20	24:00	
15.	21 January	18:00	-	22 January	13:00	
	21	18:00	-	22	18:00	
16.	24 January	24:00	-	25 January	18:00	
	24	18:00	-	25	21:00	
17.	29 January	21:00	-	30 January	3:00	
	29	18:00	-	30	5:00	
18.	10 February	17:00	-	11 February	6:00	
	10	12:00	-	11	10:00	
19.	3 March	23:00	-	4 March	7:00	
	3	18:00	-	4	9:00	
20.	31 March	10:00	-	31 March	19:00	
	31	1:00	-	31	24:00	

Table 6.2. Maximum gusts ($m\ s^{-1}$) recorded at the 7 anemometer stations during the 20 wind storms

Storm date	C-1	Gold Hill	Jones	Southern Hills	30th Street	Kent	76th Street
7 January 1969	33	45+	45+	N.O.	45+	N.O.	37
29-30 January	24	29	34	39	29	26	16
31 January	38	35	34	45+	39	27	21
19 March	26	27	33	42	36	36	26
7-8 April	32	31	33	40	27	27	18
25-26 January 1970	20	28	39	39	38	31	21
3-4 February	33	37	45+	45+	30	26	19
17 February	19	22	42	39	33	31	25
21 November	40	42	29	40	31	29	29
25 November	24	33	38	41	31	29	26
28-29 November	26	20	36	38	27	24	23
30 Nov. - 1 Dec.	26	34	40	43	41	33	26
31 December	32	26	36	36	33	29	22
20 January 1971	34	29	33	40	33	31	22
22 January	28	30	30	37	25	26	21
25 January	36	32	39	43	35	30	23
29-30 January	38	28	34	37	29	25	24
10-11 February	26	29	33	31	25	22	14
4 March	18	20	31	39	38	N.O.	23
31 March	22	26	28	41	33	29	26

45+ = exceeding the maximum of the recording chart (100 mph)
N.O. = station not in operation

CHAPTER III

RESULTS

7. Frequency Distributions

From an examination of the files of the local newspaper and the U.S. Weather Bureau storm damage records as far back as 1906, Julian and Julian (1969) obtained estimates of the annual and diurnal variability of wind storms with speeds exceeding about 30 m s^{-1} and /or causing damage in Boulder. (A similar survey of local newspapers going back as far as 1869 has very recently been done by Whiteman and Whiteman (1973, pers. comm.).) Julian and Julian's findings are shown in Table 7.1 together with the annual frequency distribution of the 20 wind storm cases used in the present study. Comparing these distributions it should be kept in mind that the 20 cases were selected from data covering only three winters and that, furthermore, the length of the winter period depended on the time of installation (early winter) and removal (late spring) of the instruments. Nevertheless, even this limited sample strongly supports Julian and Julian's conclusion that January is the prime month for wind storms in Boulder. It is also interesting to note the distribution of the 20 cases over the three winters which is 5, 3 and 12. Such large variations in the number of wind storms from year to year is likewise in agreement with Julian and Julian's findings.

Regarding the diurnal variability, an examination of the computer plots of wind traces for the 20 storm cases presented in Section 9 (Figs. 9.1 to 9.20) suggests a nighttime maximum at the Boulder stations. This, too, would agree with Julian and Julian who found the most frequent occurrence of wind storms between early evening and early morning. But there is also another type of

Table 7.1. Annual and diurnal variations of wind storms in Boulder

Annual	Month	Distribution according to Julian and Julian (1969)		Distribution of the cases used in the present study	
	September	3	(4%)		
	October	4	(5%)		
	November	16	(21%)	3	(15%)
	December	10	(13%)	2	(10%)
	January	22	(29%)	8	(40%)
	February	5	(7%)	3	(15%)
	March	11	(15%)	3	(15%)
	April	4	(5%)	1	(5%)
	May	1	(1%)		
Diurnal Time (LST)					
	0000 - 0600	32	(36%)		
	0600 - 1200	13	(15%)		
	1200 - 1800	19	(22%)		
	1800 - 0000	24	(27%)		

diurnal control indicated by the tendency of the 20 storms to either begin or end at about the time of sunrise or sunset; and temperature characteristics of the storms (defined in Section 8) do not appear to be a factor: one of the strongest warm storms, that of 7 January 1969, as well as one of the strongest cold storms, the 30 November-1 December 1970 case, began at about the time of sunset; on the other hand, the cold wind storm of 19 March 1969 and the warm storm of 3-4 February 1970 began at about the time of sunrise.

For a more objective analysis, diurnal frequency distributions were computed for the network of 7 All-Purpose anemometer stations in and around Boulder for which 5-minute data was abstracted (these are all the same type of instrument and thus comparable, except for anemometer height). The number of occurrences of 5-minute mean speeds ≥ 15 m s^{-1} and gusts ≥ 25 m s^{-1} for each 5 minute interval was obtained for the frequency distributions. Five-minute speeds rather than west wind components were used because of some missing direction data for the mean wind which in some cases would have resulted in the elimination of complete storms at a station and, because of the relatively small number of storms, probably affected the diurnal frequency distribution significantly. Also the total gust rather than its westerly component was used because of the discontinuous (once a minute) wind direction records. The effect of these limitations can be disregarded here since only the shape of the distribution of high winds is considered. To smoothen the somewhat irregular curves, total frequencies for 15 minute intervals are presented in Figs. 7.1 to 7.7. They show that at the Boulder stations (Jones, Southern Hills and 30th Street) strong winds and high gusts occurred most frequently between about midnight and 1000 MST. It is interesting to note that although Julian and Julian did not emphazise the fact, their distribution, too, has a larger number of occurrences after midnight than before. The broad maximum between midnight and 1000 consists however of two peaks, one at about 0100 and another at 0700 with a marked minimum between 0500

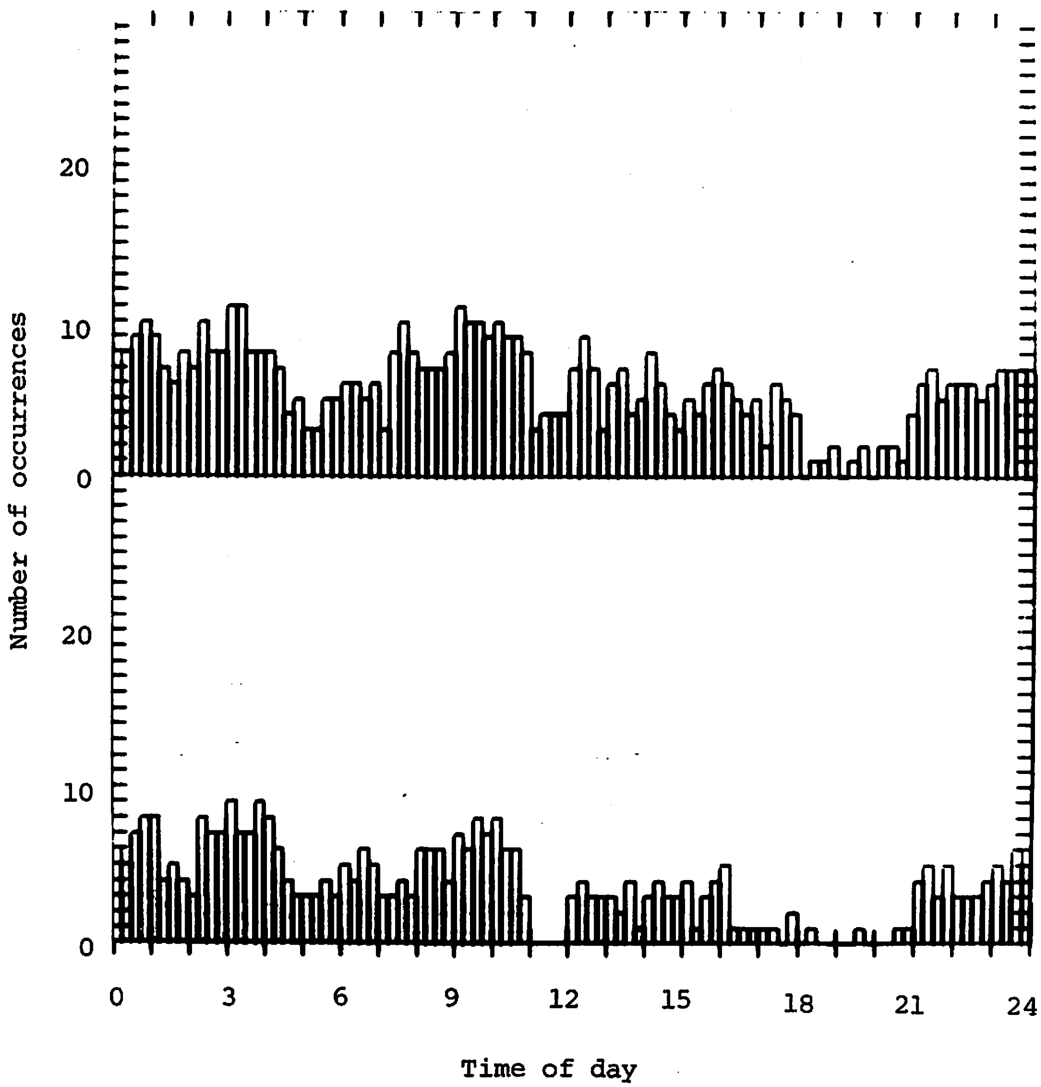


Fig. 7.1. Total number of occurrences of 5-minute gust speeds $\geq$25 m s^{-1} (top) and 5-minute mean speeds $\geq$15 m s^{-1} (bottom) by 15 minute intervals for 20 Boulder wind storms, at C-1 station.

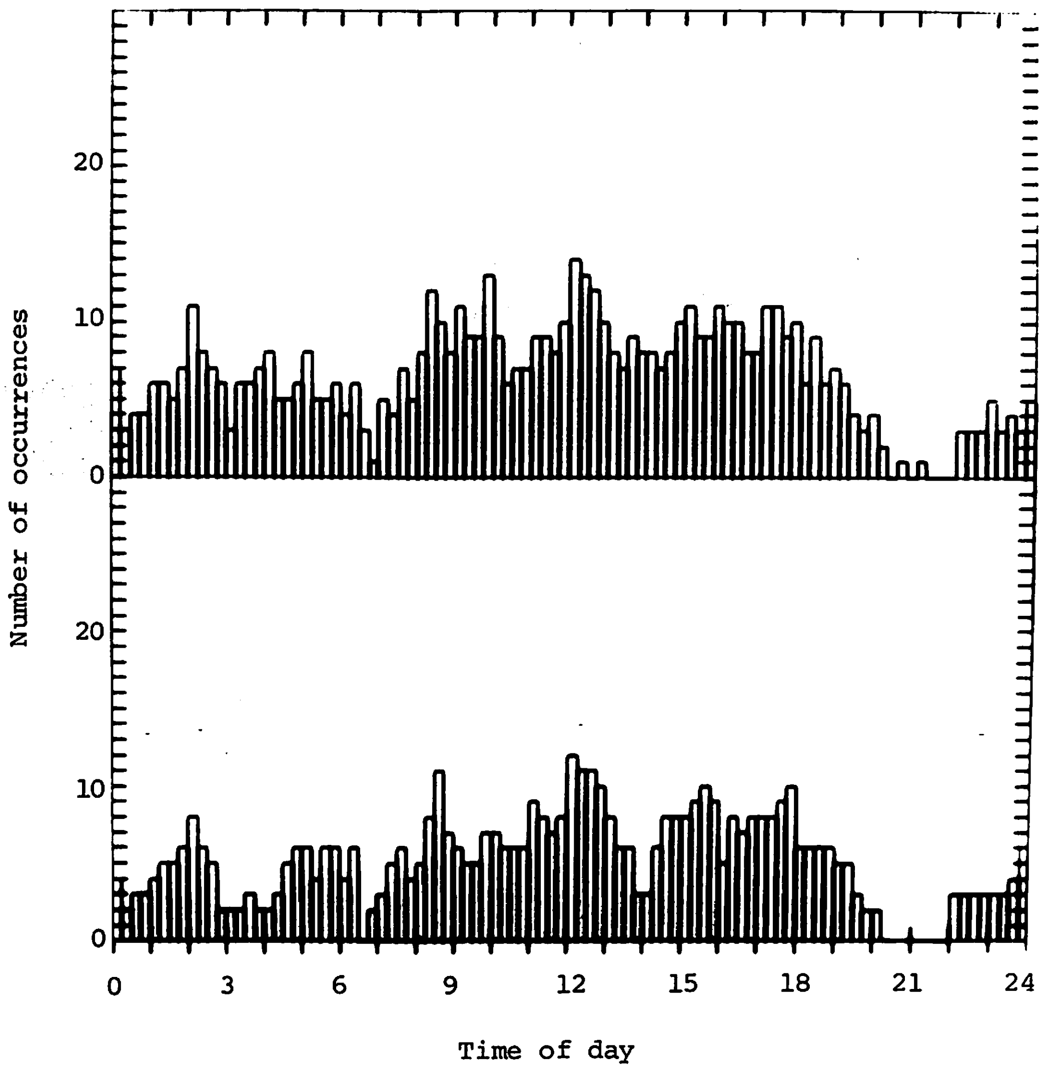


Fig. 7.2. Total number of occurrences of 5-minute gust speeds ≥ 25 m s^{-1} (top) and 5-minute mean speeds ≥ 15 m s^{-1} (bottom) by 15 minute intervals for 20 Boulder wind storms, at Gold Hill station.

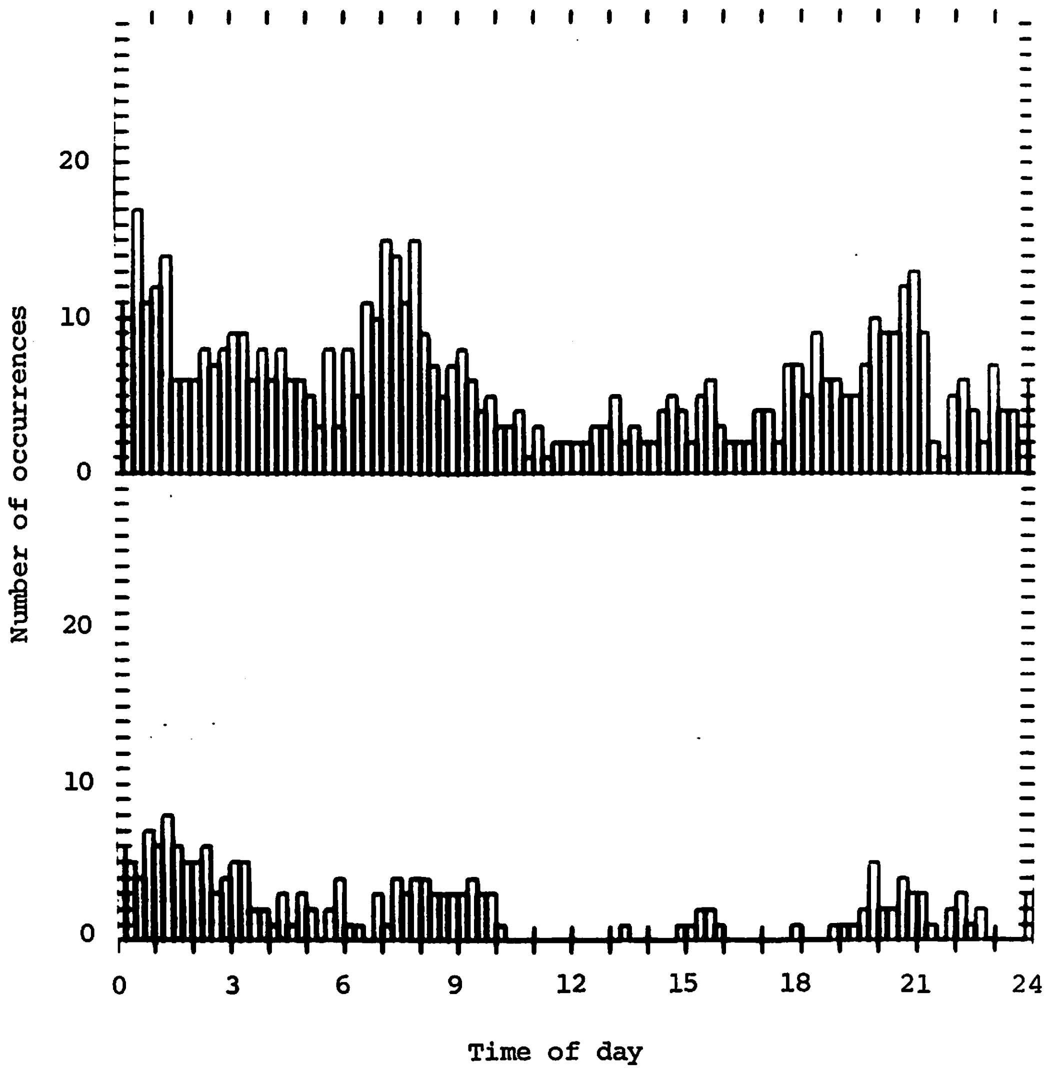


Fig. 7.3. Total number of occurrences of 5-minute gust speeds ≥ 25 m s^{-1} (top) and 5-minute mean speeds ≥ 15 m s^{-1} (bottom) by 15 minute intervals for 20 Boulder wind storms, at Jones station.

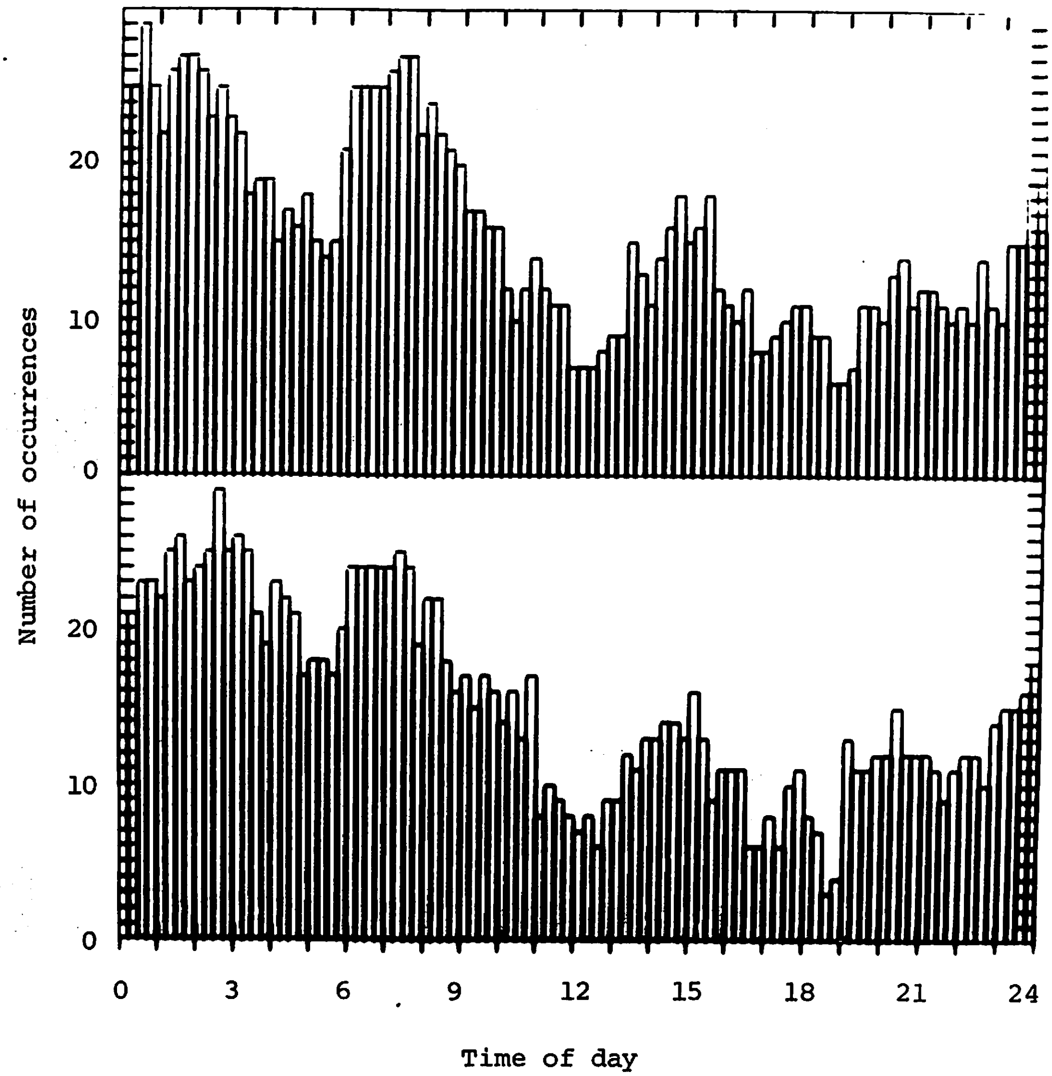


Fig. 7.4. Total number of occurrences of 5-minute gust speeds ≥ 25 m s^{-1} (top) and 5-minute mean speeds ≥ 15 m s^{-1} (bottom) by 15 minute intervals for 20 Boulder wind storms, at Southern Hills station.

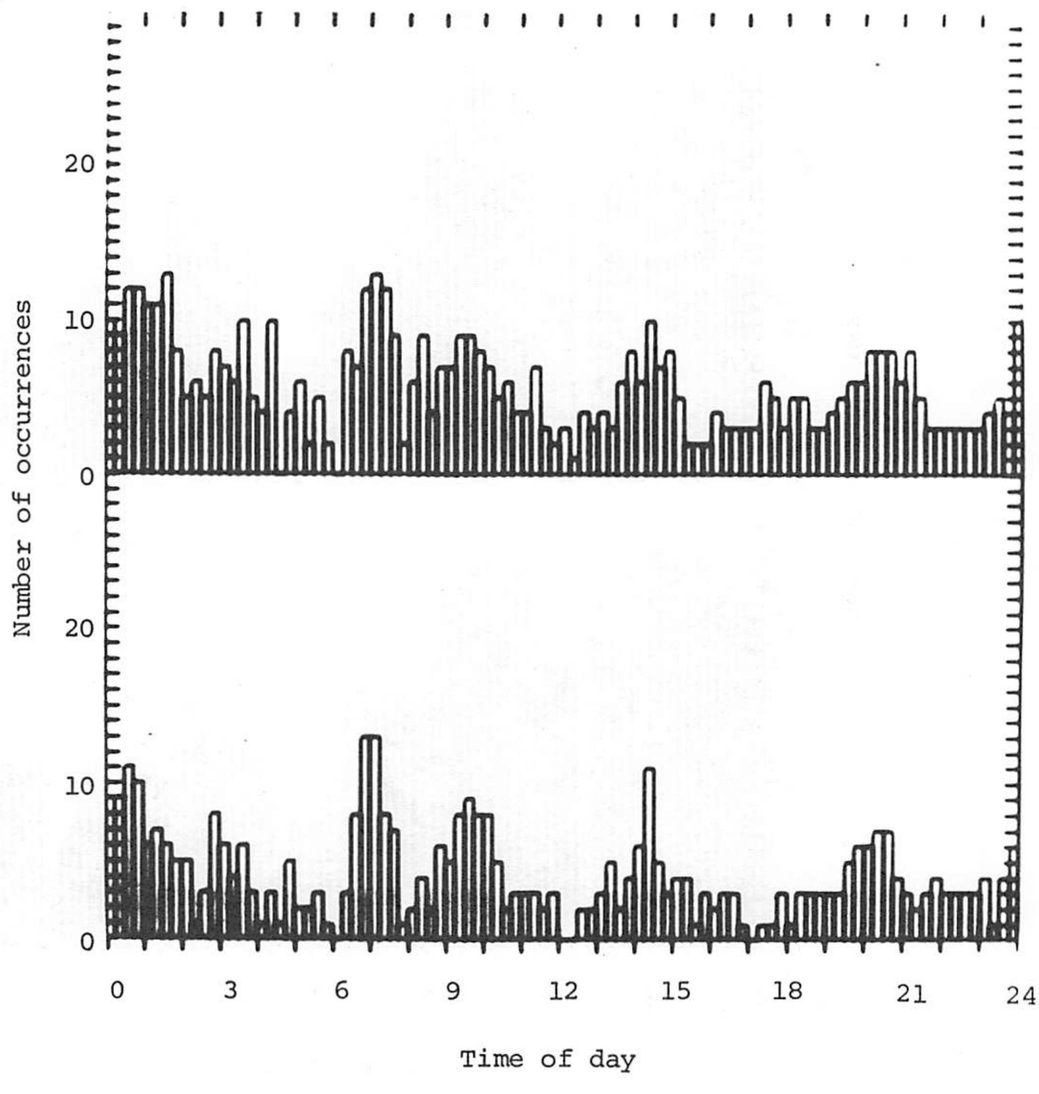


Fig. 7.5. Total number of occurrences of 5-minute gust speeds ≥ 25 m s^{-1} (top) and 5-minute mean speeds ≥ 15 m s^{-1} (bottom) by 15 minute intervals for 20 Boulder wind storms, at 30th Street station.

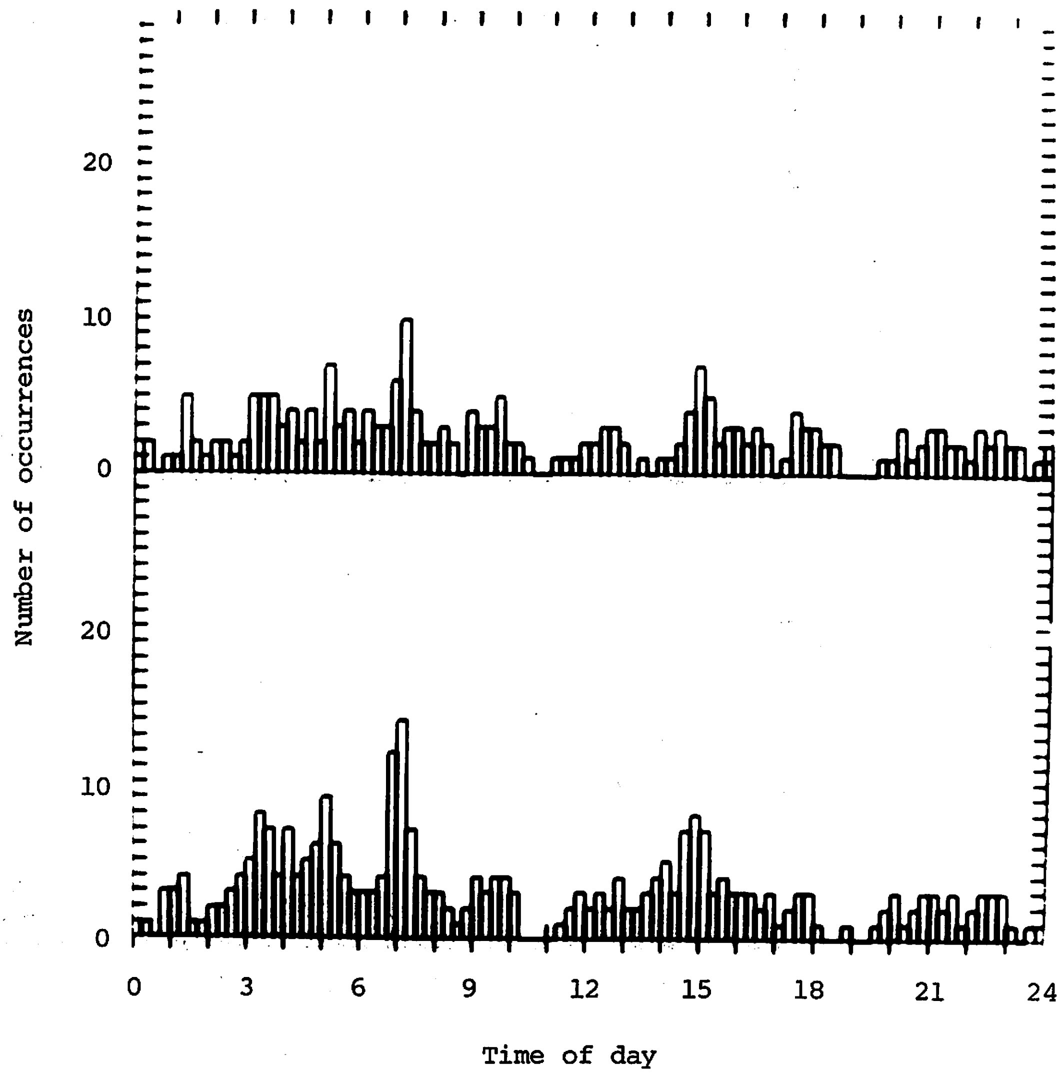


Fig. 7.6. Total number of occurrences of 5-minute gust speeds ≥ 25 m s^{-1} (top) and 5-minute mean speeds ≥ 15 m s^{-1} (bottom) by 15 minute intervals for 20 Boulder wind storms, at Kent station.

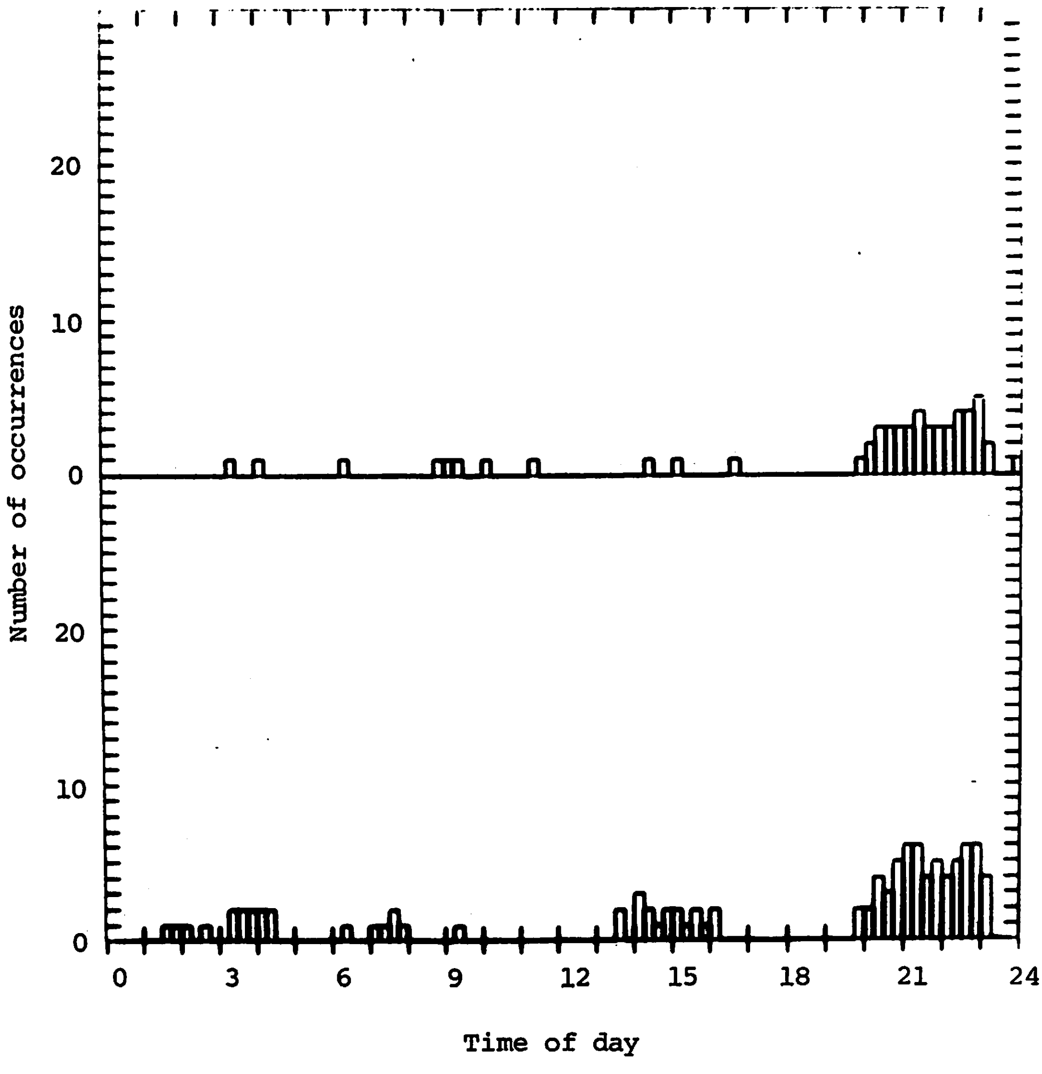


Fig. 7.7. Total number of occurrences of 5-minute gust speeds ≥ 25 m s^{-1} (top) and 5-minute mean speeds ≥ 15 m s^{-1} (bottom) by 15 minute intervals for 20 Boulder wind storms, at 76th Street station.

and 0600. There are also minor peaks around 1500 and 2100, separated by a minimum around 1700-1800. Since the time of sunrise and sunset during the winter months at Denver are 0630-0730 and 1630-1730 respectively, the frequency minima at these times must be the result of the already mentioned tendency for the storms to either begin or end at about the time of sunrise or sunset.

At the Kent station (about 6 km to the east of Boulder) there is no indication of the 0100 peak in the frequency distribution but rather a broad maximum between about 0300 and 0700, and the 1500 peak is of almost equal importance. At the 76th Street station (about 9 km to the east) only the 2000-2100 peak is outstanding. Since wind speeds at the 76th Street station are generally much lower (see Section 9), its frequency distribution is obviously the result of a few very strong winds. Closer examination shows that only 4 storms and one additional phase contributed to it. One of them occurred in the early morning hours, one around sunrise, another in the afternoon and two in the evening (the warm storm of 7 January 1969 and the first phase of the cold 30 November-1 December 1970 case), which seems to suggest that neither their temperature characteristics nor the time of their occurrence are determining factors in the eastward extent of Boulder's storms as has been assumed. The apparent shift in the diurnal frequency maximum between Boulder and the stations to the east may therefore be the result of the very small sample of strong winds and hence not be significant.

The frequency distributions for the mountain stations, Gold Hill (about 13 km west of and 980 m above Boulder) and C-1 (about 23 km west of and 1450 m above Boulder) do not represent diurnal variations of high winds at these stations per se but only for periods with wind storms in Boulder. There are some similarities between the distributions for the Boulder and slope stations, particularly C-1, but the broad daytime maximum, most noticeable at Gold Hill, and the marked minimum about 2000-2100, are out-of-phase with the Boulder stations. This agrees with the results of the following Section 9 which show that the wind maxima on the slopes do not tend to coincide with storms in Boulder.

Because of the known preference of wind storms in Boulder for the night hours, cooling in the boundary layer has been suggested as a possible causative factor (Vergeiner 1971; Lilly and Zipser 1972). Although temperature differences between the slope and Boulder stations at the height of the storms were found to be close to zero (Section 8) radiative cooling could possibly be a trigger mechanism considering the high frequency of storms in the early morning hours when the mountain wind is best developed. On the other hand, the apparent relationship between sunrise/sunset and onset/cessation of storms in Boulder suggests another kind of diurnal control. It may be that major changes in the radiation balance can significantly affect critical atmospheric variables by shifting them into the region either favourable or unfavourable for wind storms in Boulder. For example, diurnal changes in the lapse rate on the windward side may be one such critical factor, to be discussed further.

8. Temperature Characteristics

Although Boulder's winds are generally assumed to belong to the class of warm downslope winds, locally known as Chinooks (for example, Harrison 1956; Beran 1967; Houghton and Kasahara 1968; Julian and Julian 1969; Sangster 1970; Vergeiner and Lilly 1970; Vergeiner 1971) their temperature characteristics have never been studied. Some work has been done on "Chinook winds" in the Colorado Front Range area but, like other studies of foehn-type winds around the world, they have been considered primarily as a temperature phenomenon and consequently temperature criteria have been used to select the study cases. Ives (1950), for instance, used a definition based on the classical foehn theory (a potential temperature which is higher in the lee, i.e. Denver, than on the windward side, at Fraser; precipitation on the windward side; and a westerly wind direction in the lee). Beran (1967) used the dew-point spread as the primary indicator of his Chinook areas.

Since the emphasis in the present study is on wind speed rather than temperature, the 20 cases were selected on the basis of wind speed criteria which permitted an objective unbiased analysis of the temperature characteristics of the wind storms. However, as was pointed out in Section 2, temperature changes that are taking place in the atmosphere may be masked to some degree at lower levels, for example by a surface inversion, which in the extreme case may result in a cold air foehn. Thus, in order to determine the temperature characteristics of Boulder's wind storms, both the genetic characteristics (determined from upper mountain slope and free air data) and apparent temperature characteristics at the surface in Boulder and Denver were studied. Because of some uncertainty in the available data, three upper level and two surface temperature parameters were selected to serve as a basis for grouping the storms according to their temperature characteristics.

An upper 'slope' temperature was obtained by averaging the hourly potential temperature for the three mountain stations B-1 (at approx. 740 mb), C-1 (at approx. 700 mb) and D-1 (at approx, 635 mb). A change in this 'slope' temperature equal to or exceeding $|1°C\ hr^{-1}|$, if sustained over a period of at least three hours, was considered to be 'marked'. If the end of such a temperature change occurred within 6 hours prior to the wind in Boulder, the total temperature change was considered to be representative of atmospheric changes related to the wind storm occurrence in Boulder. These arbitrary criteria were decided upon after examination of typical diurnal changes and consideration of the problem of associating a wind storm with a temperature change more than 6 hours prior to it.

Over half of the wind storms were found to follow the classical example of downslope winds (Section 2), with a temperature change (increase or decrease) occurring in association with the onset of the wind. An average of 4 hours (ranging from 0 to 6 hours) elapsed between the end of the temperature change and the beginning of the wind. In 8 cases, however, there was no marked temperature change within 6 hours prior to the onset of the storm. These were considered to be indifferent cases.

There was generally little difference in potential temperature between Boulder and the upper slopes as soon as the winds became westerly in Boulder. The temperature also remained remarkably constant at these stations; the lack of diurnal variations for the duration of the wind is due to the strong mixing of the air. Denver's temperature, however, showed marked fluctuations because of generally lower wind speeds there, which frequently permitted the reestablishment of the surface inversion at night, and radiative warming during the day. Another typical characteristic was found to be the occurrence of 'pauses' in wind speed which separated 3 of the 20 storm cases into two distinct phases of wind. (The short wind speed peaks associated with frontal passages were not considered.) In one case a pause was found to have separated two phases each preceded by a temperature change, the second phase was therefore considered a separate case.

In some instances, the temperature change was found to be somewhat similar in sign and time of occurrence to that expected from normal diurnal variations (the average diurnal range for the three mountain stations during the winter is approximately 9°C) and the wind speed at these stations during the period of temperature change was not always high enough to dismiss this factor. However, the temperature of the free air at the 700 mb level is relatively unaffected by diurnal influences, varying only about ± 2°C (Denver Forecast Office 1972, pers. comm.), except for frontal passages. To confirm the validity of the 'slope' temperature change, changes in a comparable layer in the 'free air' over Denver (approx. 750-650 mb) were analysed.

Because of the sub-synoptic nature of the winds it is extremely difficult to make use of synoptic data. Nevertheless, an effort was made to interpret the available data meaningfully. To determine the temperature change in the free air, a sounding prior to and one during the wind were used if possible. In a number of cases the temperature change occurred at about the time of the sounding, and a sequence of sounding profiles had to be compared to determine which represented the state of the atmosphere before, during, and after the air mass change (if any) and the appropriate soundings were then used to determine the change in temperature. It also proved to be somewhat difficult to determine the free air temperature when the lapse rate was not adiabatic. For example, there is a tendency for an inversion to be located at about the upper part of the 750-650 mb layer during wind storms, as will be discussed later. If this was the case, only the temperature below the level of the inversion was considered, assuming that the air involved in the storm had been that below the inversion. Fig. 8.1 shows the close relationship obtained between the values from the upper slope and the free air, despite the many difficulties of interpretation. Comparison between the regression line and the line of equal change (F-tests of slope and elevation) showed no differences between them. This supports the view that the marked temperature changes observed at the upper slopes represent real air mass changes in the atmosphere. (The clustering along the 0°C slope temperature line is the result of the definition used according to which a change of less than 3°C at the upper slopes was not noted.)

Horizontal advection was selected as the third means of determining the generic temperature characteristics of the wind

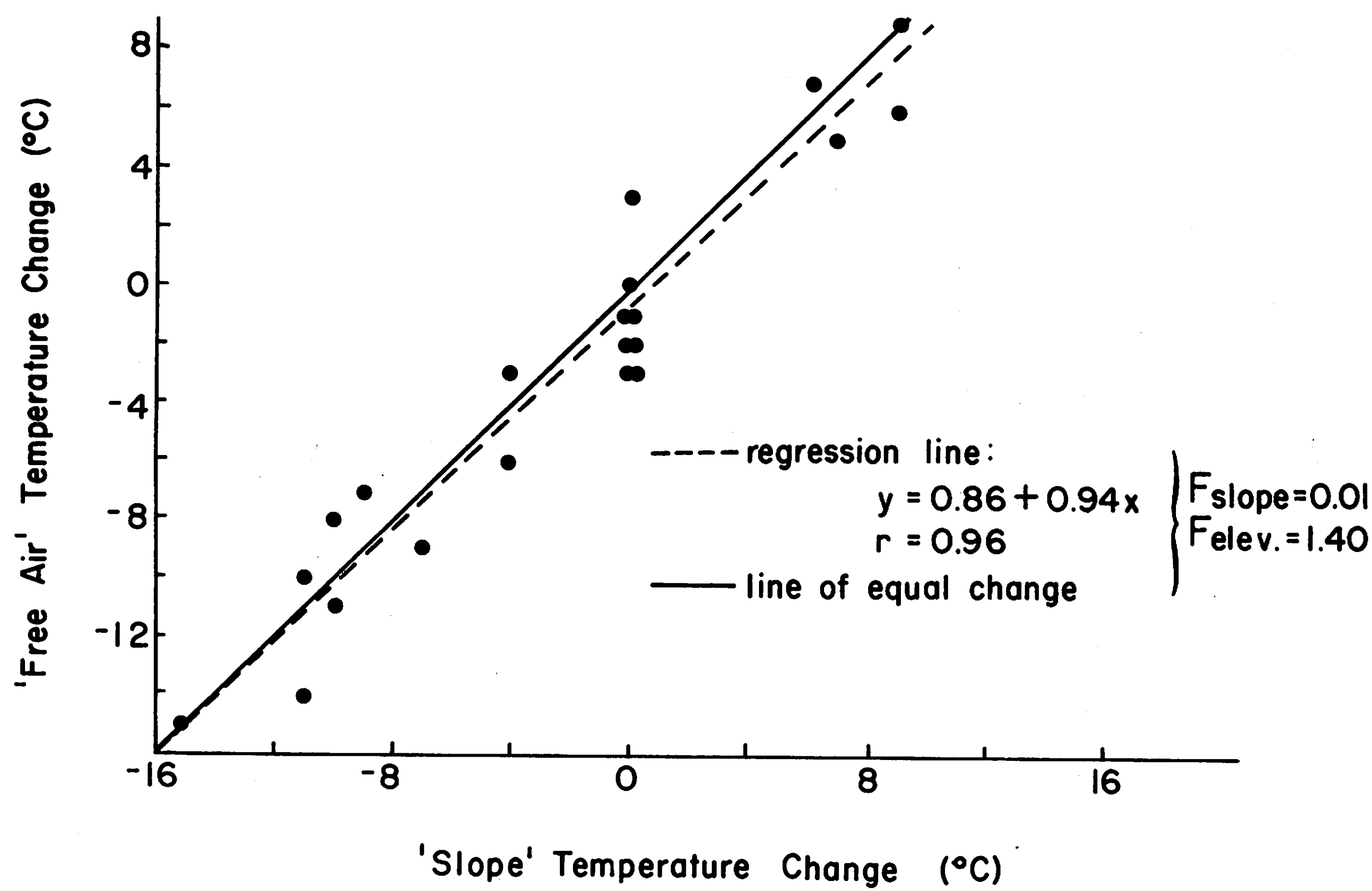


Fig. 8.1. Scatter diagram relating the free air temperature change between about 635 mb and 740 mb, as obtained from a sequence of soundings for Denver, to the surface temperature change observed at the three upper mountain stations B-1 (740 mb), C-1 (700 mb) and D-1 (635 mb).

storms. The advection term was calculated from

$$-v_s \frac{\partial T}{\partial s}$$

where T = potential temperature
v_s = wind component normal to the horizontal temperature gradient
s = distance.

The temperature gradient over 4 degrees of latitude or approximately 440 km centered on Denver (an arbitrary choice which seemed appropriate for the spacing of the rawinsonde stations and the scale of the analysis) and the direction of this gradient were determined from the 700 mb constant pressure charts. The wind speed and direction for that level over Denver were obtained from the soundings. In general, the constant pressure charts and soundings closest to the time of temperature change and onset of the storm were used to calculate the horizontal advection. Fig. 8.2 compares the results of this method to the 'slope' temperature changes discussed above and shows the good agreement between the two parameters; decreases in temperature at the upper mountain slopes are associated with cold air advection and increases with warm advection. The slope of the regression line in Fig. 8.2 suggests that the temperature change observed at the slopes occurred over a period of about 16 hours. The actual time over which the change took place was, however, generally less than half this value which points out that the horizontal advection values in Fig. 8.2 are, unavoidably, averages over the distance of 440 km used for the calculation but that somewhere along this distance there usually was a frontal zone.

To determine the genetic temperature characteristics of the storms, the results presented in Figs. 8.1 and 8.2 were used as a

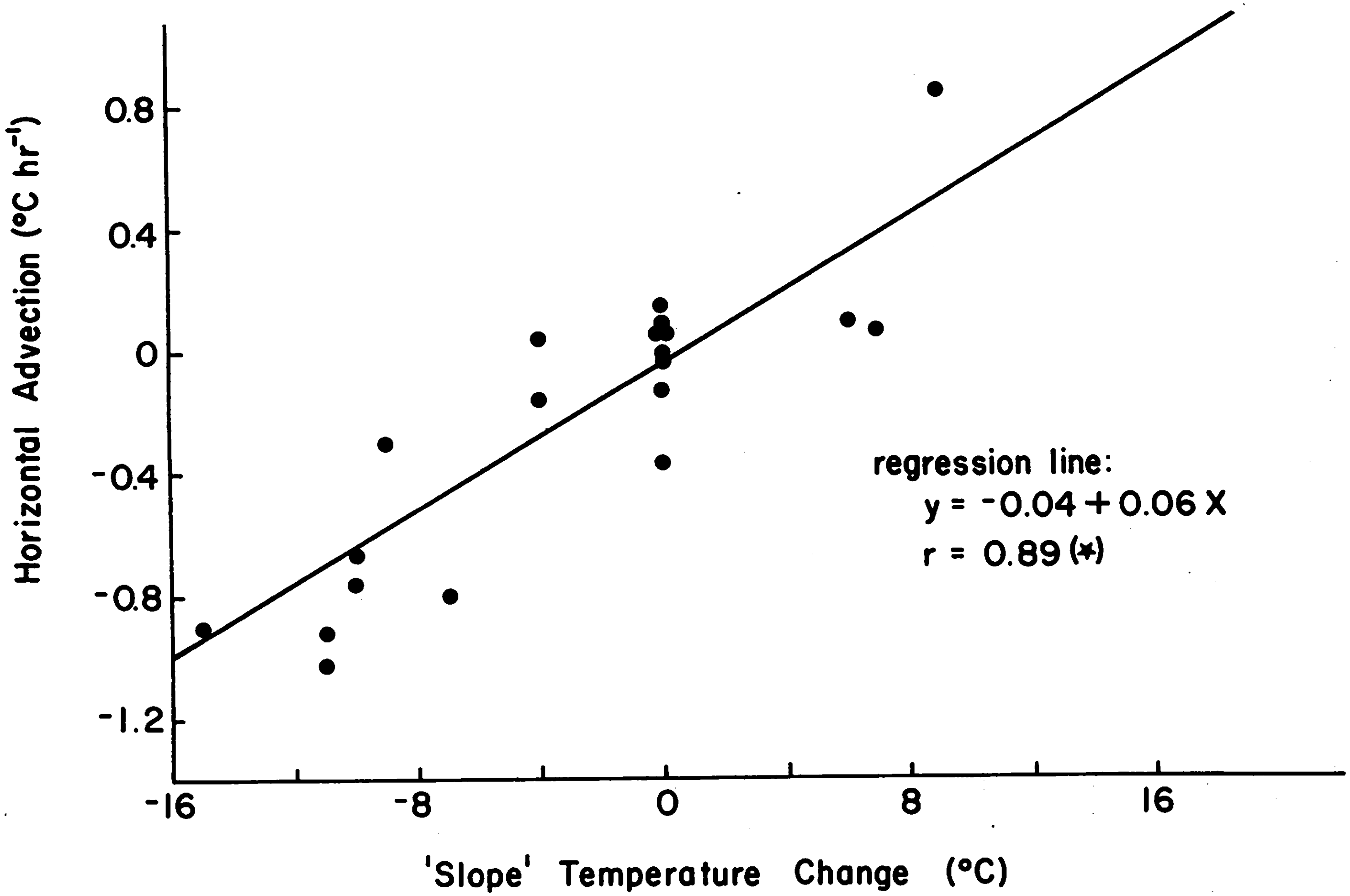


Fig. 8.2. Scatter diagram relating horizontal advection at the 700 mb level over Denver to the change in surface temperature observed at the three mountain stations B-1 (740 mb), C-1 (700 mb) and D-1 (635 mb).

basis for grouping them. This is summarized in Table 8.1; a cold or warm storm had to satisfy at least two of the three criteria: change in slope temperature $\geq |4°C|$; change in free air temperature $\geq |4°C|$; horizontal advection $\geq |0.2°C\ hr^{-1}|$. That some of the cases were more difficult to fit into this simple scheme than others only shows that the whole range of wind types is possible and that the grouping is only an aid in understanding the general processes behind the complexity of observations. The important conclusion to be drawn from the table is that <u>Boulder's wind storms are not all warm downslope winds, as has generally been assumed. About one-half of the cases studied were found to be genetically cold (associated with decreasing temperatures in the free atmosphere) while less than one-quarter were warm.</u>

Table 8.1. Classification of wind storms according to upper slope and free air temperature characteristics

Description	'Slope' temperature change (°C)	'Free air' temperature change (°C)	Horizontal advection for 700 mb level (°C hr^{-1})	Number of cases
Cold	≥ -4	≥ -4	≥ -0.2	9
Indifferent	$-4 > \Delta T < +4$	$-4 > \Delta T < +4$	$-0.2 > \frac{\Delta T}{\Delta t} < +0.2$	8
Warm	$\geq +4$	$\geq +4$	$\geq +0.2$	4

To determine the apparent temperature changes at the foot of the mountains, which have always been the basis of conventional definitions of warm and cold downslope winds around the world (as discussed in Section 2), 'marked' cooling or warming (defined as for the 'slope' temperature change) just prior to or during strong winds in Boulder and associated wind maxima in Denver were noted. (Wind storms in Boulder do not necessarily mean strong winds in Denver or simultaneous wind maxima, as will be discussed in Section 9.) These temperature changes were not always simultaneous or equal to those determined for the upper slopes because of time lag and differences in elevation. The time of occurrence of the beginning of the temperature change at lower levels is obviously important since the removal of shallow surface inversion layers by the wind at night and already existing high temperatures during the day would strongly modify such temperature changes.

Figs. 8.3 and 8.4 relate temperature changes at Boulder and Denver to those at the upper slopes (the upper slope temperature changes are taken to represent the true changes since their close relationship with the free air changes has been shown above and continuous temperature records make them superior to the 12-hourly upper air data). Statistical test of the regression lines in Figs. 8.3 and 8.4 show that the temperature changes in the atmosphere and those at the foot of the mountains in Boulder and Denver are highly correlated although there is also a marked tendency toward apparent warming at lower levels (i.e., a significant elevation of the regression line above the line of equal change, but with similar slope). In some instances, therefore, there is a

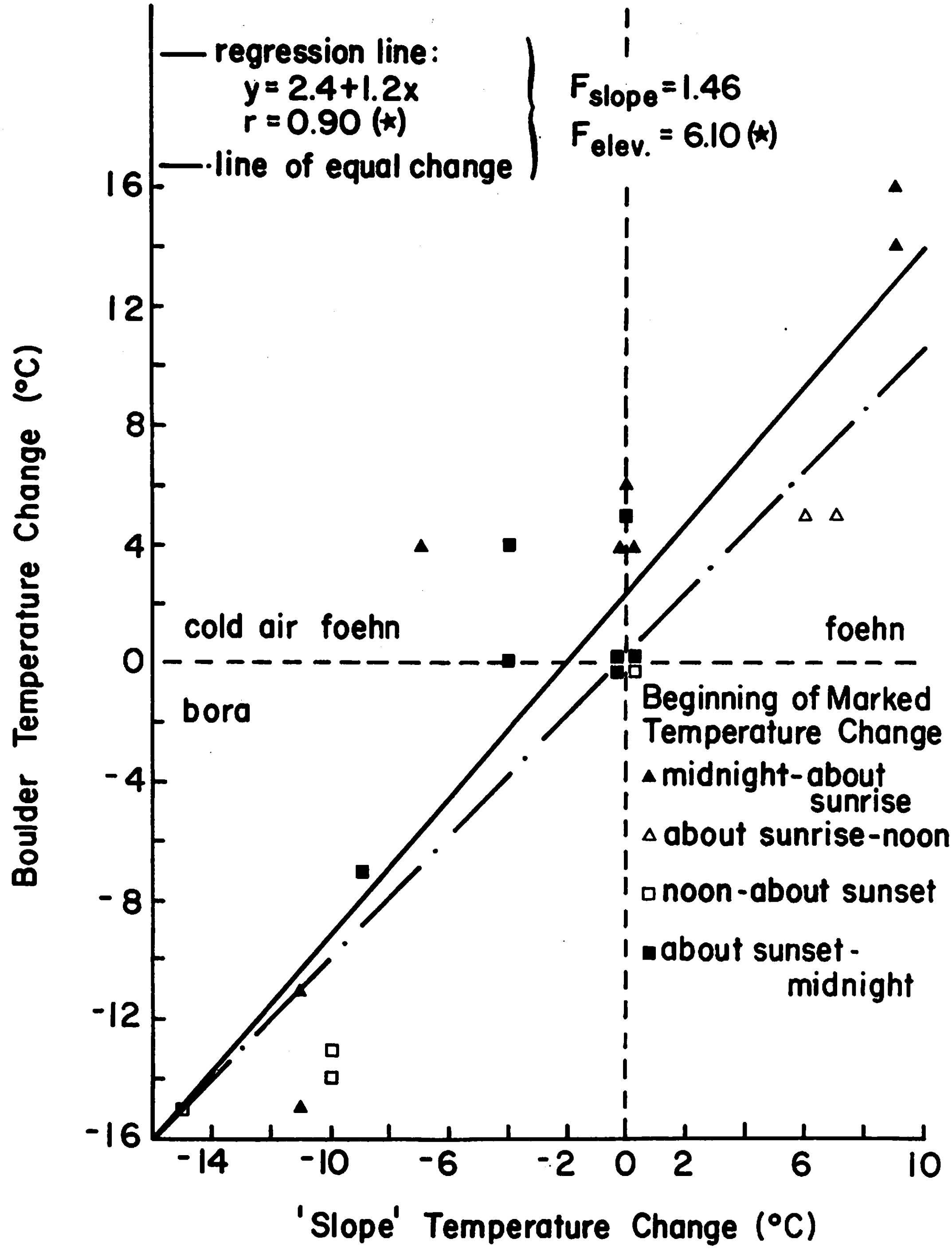


Fig. 8.3. Scatter diagram relating the temperature change observed in Boulder (in association with a wind storm there) to that at the upper mountain stations.

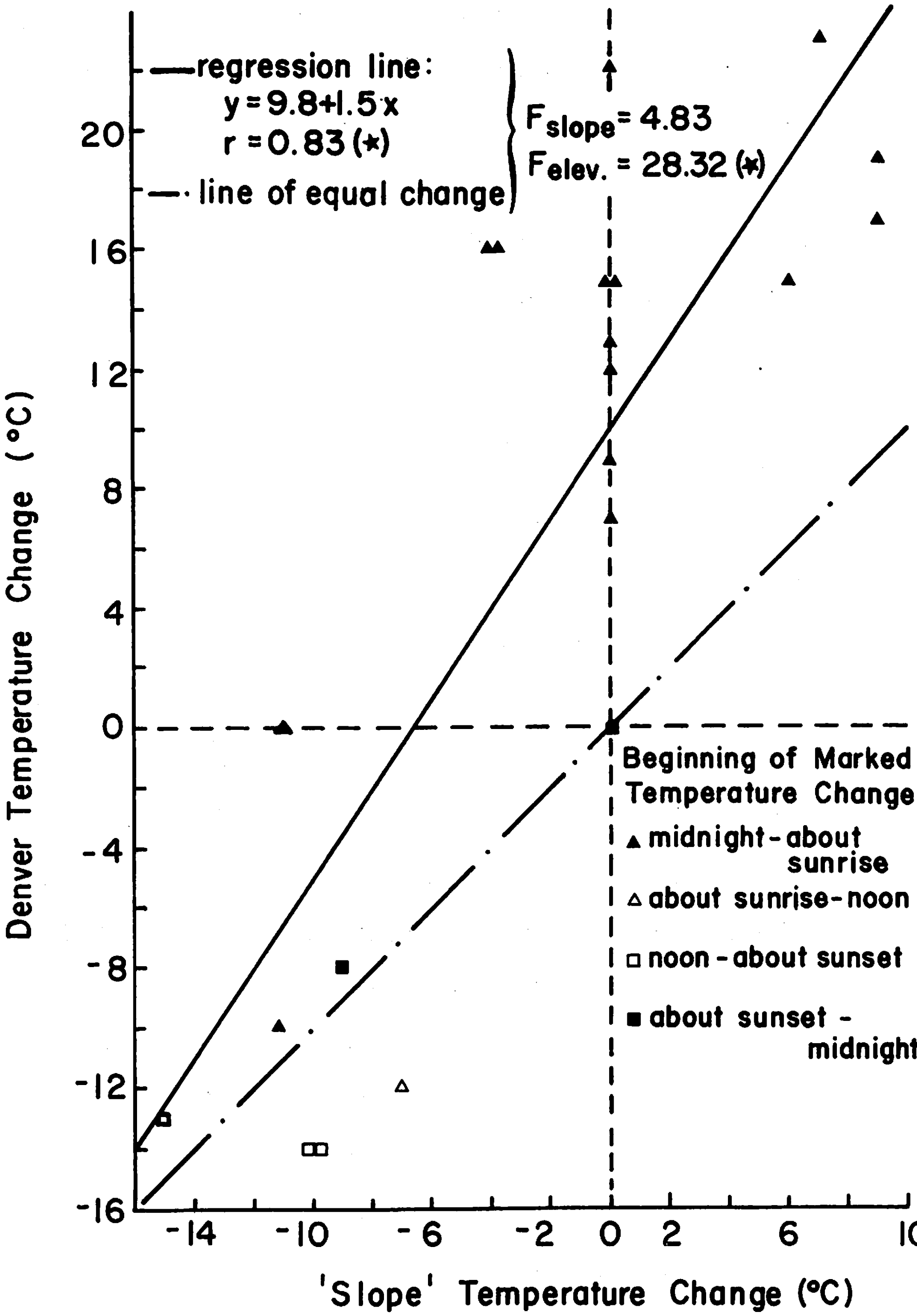


Fig. 8.4. Scatter diagram relating the temperature change observed in Denver (in association with a wind maximum there and a wind storm in Boulder) to that at the upper mountain stations.

shift in the temperature change associated with the wind storms from cooling at upper levels to warming below, thus producing the whole range of possible wind types, from bora to cold air foehn to true foehn. The largest warming at lower levels, as compared with the upper slopes, tends to be associated with a storm occurring in the early morning hours when the surface inversion is strongest. This warming is more pronounced in Denver as compared with Boulder, as expressed in the significantly higher elevation of the Denver regression line above the line of equal change as compared with Boulder. The most outstanding feature of Figs. 8.3 and 8.4, however, is the fact that some of the wind storms are extremely cold, causing decreases in temperature at the upper slopes as well as Boulder and Denver by up to 15°C, while the extremely warm cases can produce temperature rises by 9°C at the upper slopes and 15°C to 20°C in Boulder and Denver. <u>Some of Boulder's wind storms can therefore be as cold as most of the bora-type winds around the world and some of them as warm as any of the foehn-type winds</u>.

Another measure of the apparent warming or cooling effect of the wind storms at lower levels is the difference between the temperature observed during the storm and the average temperature for the time of day and season. Using normal daily maximum and minimum temperature (for the period 1931 to 1960) and taking into consideration times of sunrise and sunset, a curve of daily temperature variations was produced for each of the winter months for Boulder and Denver. The actual temperature at the time of the wind maximum at the two cities (averaged over three hours and in the case of Boulder over all available stations) was compared with the expected temperature as given by the diurnal curve for the month. Figs. 8.5 and 8.6 show that there is little correlation between temperature changes in the atmosphere and temperature deviations from the average diurnal curve at the foot of the mountains, but that there is a mean positive departure from the diurnal trend of 8°C for both Boulder and Denver. For Boulder the mean nighttime temperature departure of +11°C is significantly higher than that for the daytime (+2°C); for Denver the difference between +11°C for the night and +5°C for the day is smaller and not significant because of diabatic heating during the day, a point to be discussed further. Only very few of the wind storms produce negative departures; most of them appear as winds with positive anomaly, with the more pronounced departures again being caused by those occurring at night.

To determine the apparent temperature characteristics of the storms in Boulder and Denver, the results presented in Figs. 8.3, 8.4, 8.5 and 8.6 were used as a basis for grouping them. This is summarized in Table 8.2; the members in each category had to satisfy both criteria. The results show that most of the wind storms appear to be warm at lower levels and that this is more pronounced in Denver than in Boulder.

In Table 8.3 are combined the results presented in Table 8.1 and 8.2. Of the 20 cases (plus one phase) of wind storms in Boulder reaching hurricane-force speeds, about one-half were found to be genetically cold and less than one-quarter warm. But, whereas the true foehn winds retained their characteristics at all levels, many of the cold and indifferent winds turned increasingly into apparently

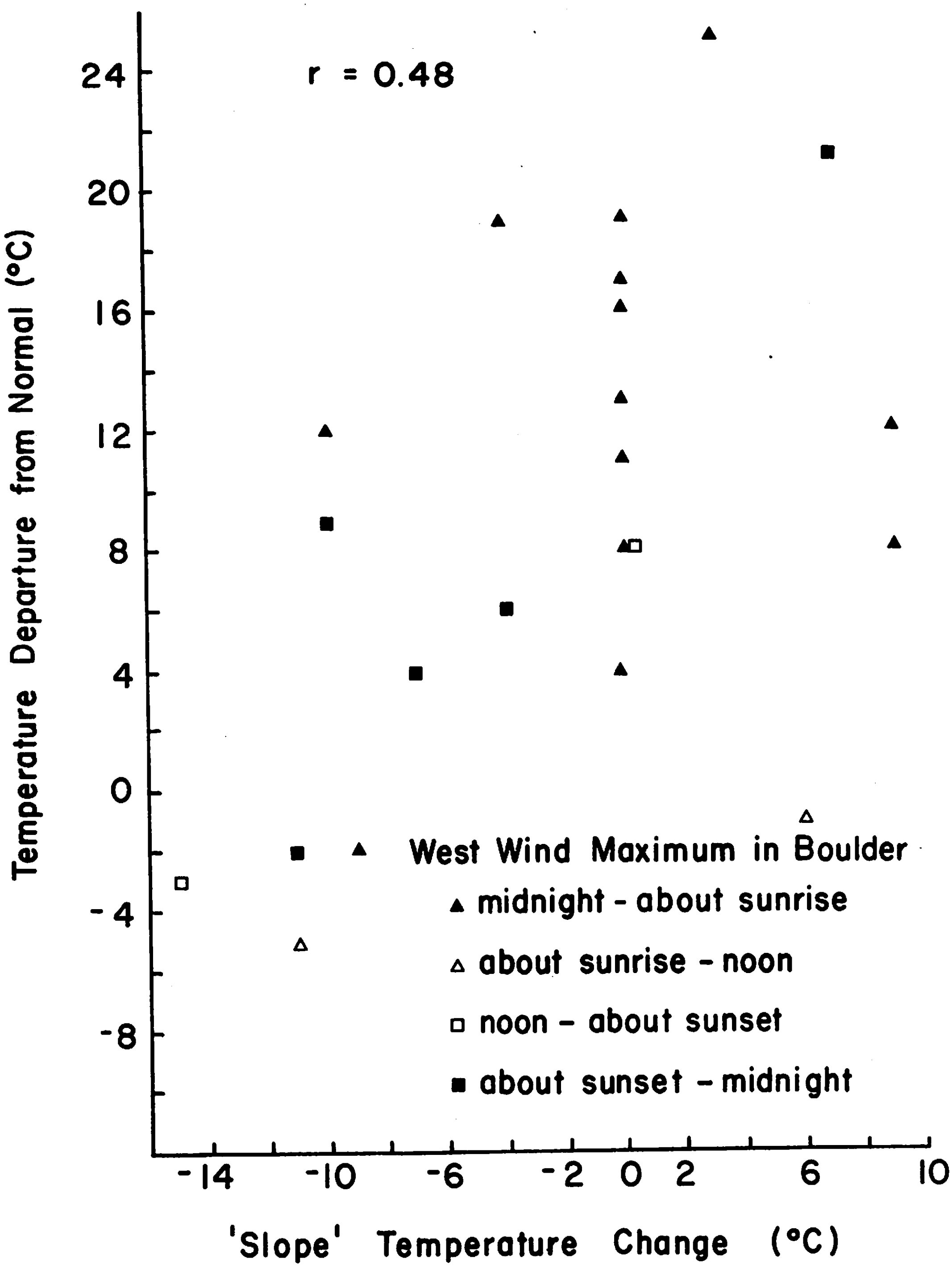


Fig. 8.5. Scatter diagram relating the departure of temperature from normal for the hour of the day and the month at Boulder at the height of the wind storm there to the change in temperature observed at the upper mountain stations.

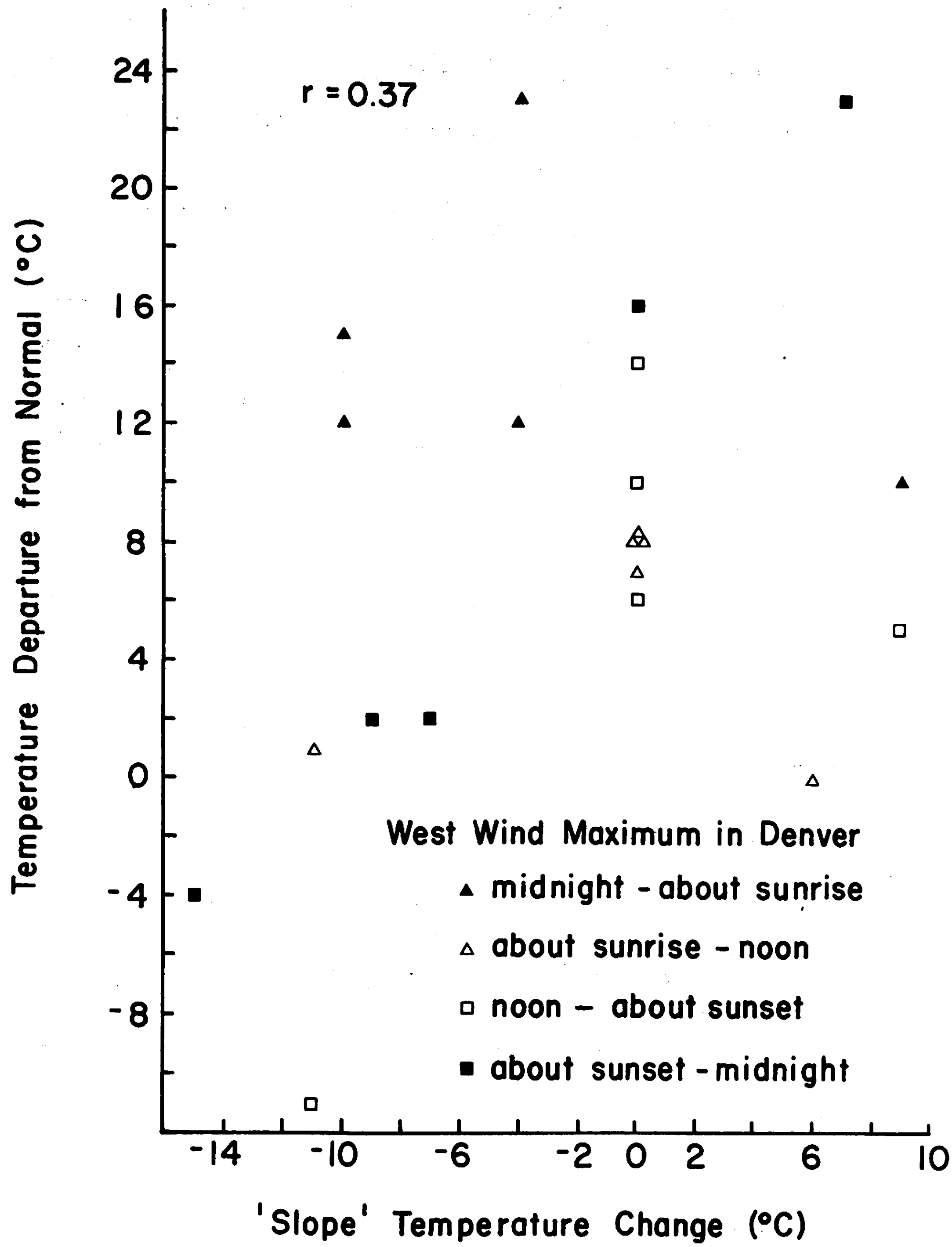


Fig. 8.6. Scatter diagram relating the departure of temperature from normal for the hour of the day and the month at Denver at the time of the wind maximum there to the change in temperature observed at the upper mountain stations.

warm winds with decreasing elevation. In Boulder less than one-quarter of the winds remained cold while three-quarter were warm and in Denver almost all appeared to be foehn-type winds.

Table 8.2. Classification of wind storms according to apparent lower level temperature characteristics

Description	Boulder or Denver temperature change	Boulder or Denver deviation from normal	Number of cases	
			Boulder	Denver
Bora	≥ -4	<0	4	2
Indifferent	$-4 > \Delta T < +4$	$-4 > \Delta T < +4$	2	0
Foehn	$\geq +4$	≥ 0	15	19

Table 8.3. Distribution of wind storm types at different elevations

Types according to genetic as well as apparent characteristics	Upper slopes /Free air	Boulder	Denver
Bora	9	4	2
Cold air foehn		5	7
Indifferent	8	2	0
Indifferent foehn		6	8
Foehn	4	4	4

The major reason for the difference between genetic and apparent temperature characteristics is the already mentioned removal or supression of a surface inversion at lower levels. In addition, there is marked evidence of diabatic processes between Boulder and Denver. Fig. 8.7 shows that, while there is no general relationship between mean hourly west wind for the 3-hour period with the strongest wind in Denver and the average Denver-Boulder temperature difference for that period, the Denver temperature was found to be raised significantly above that of Boulder by 3°C whenever the wind maximum occurred in Denver during the daytime hours, and by 1°C at night. Some of the warming effect of the wind in Denver is, therefore, the result of diabatic heating which may be especially large in wind storm conditions because the sky is usually clear and any snow on the ground in the plains would be quickly removed. Diabatic warming may at least be partly responsible for the larger number of apparent foehns at Denver as compared with Boulder. Along the lee slopes diabatic processes such as radiative cooling and evaporation of snow are also a factor to be considered and have even been suggested to represent part of the descent mechanism of the wind (for example, Berg 1952; Lilly and Zipser 1972). However, the difference in temperature between the slopes and the free air at the same level over Denver which may be an indication of such processes was found to be -5°C prior to the wind and -4°C during storms in Boulder, and the difference of 1°C is not significant. The mixing ratio (calculated from equations given in Section 4) and its change between stations should have made it possible to estimate the cooling due to evaporation. But because

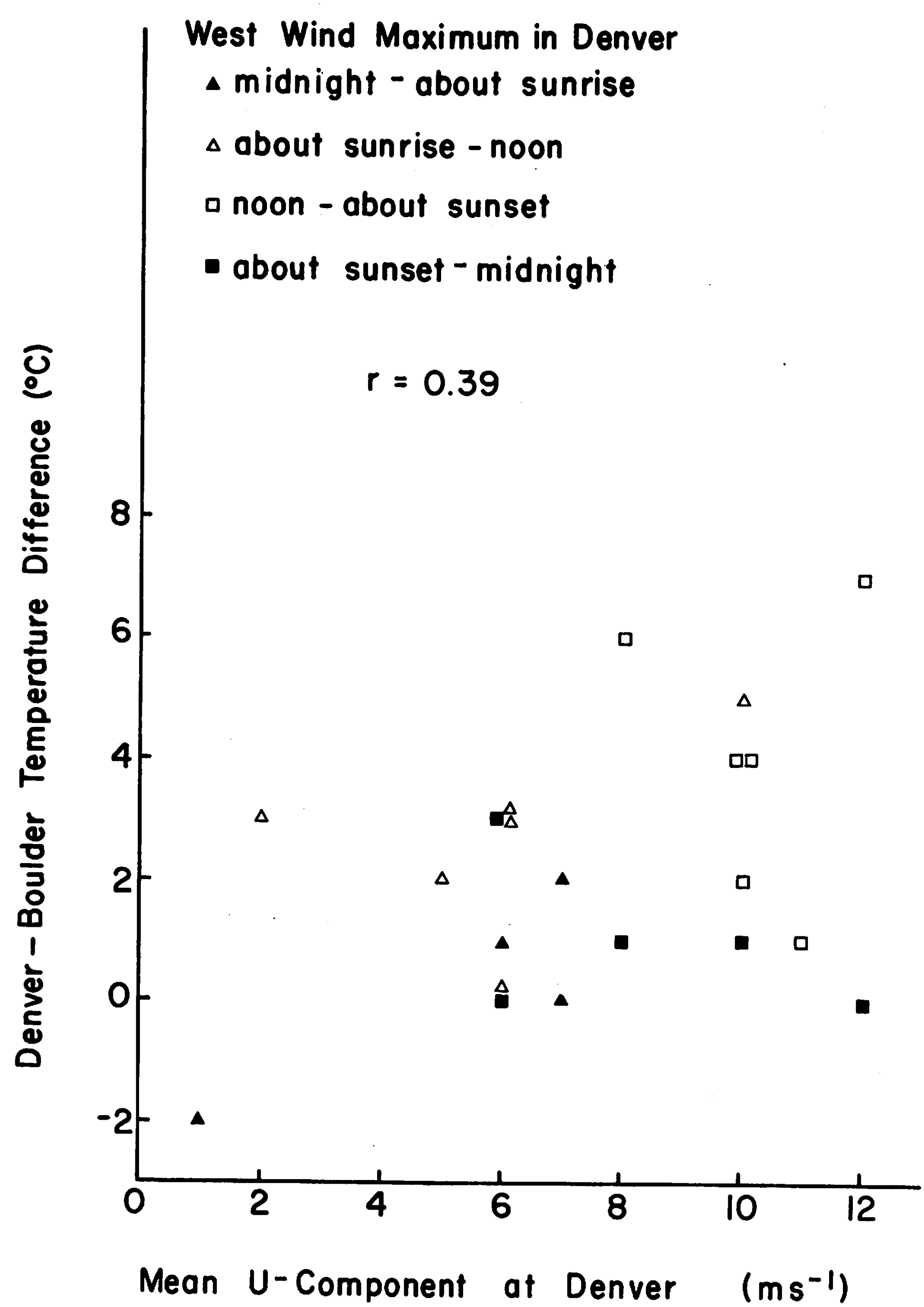


Fig. 8.7. Scatter diagram relating the difference in temperature between Denver and Boulder at the time of the wind maximum in Denver to the average westerly surface wind component at Denver for the same period.

of the sensitivity of the calculated mixing ratio to the accuracy of the humidity measurements (which are difficult to make at higher elevations and low temperatures with the conventional hair hygrometer) the calculated mixing ratio was found to be too unreliable to be used. The amount of radiative cooling to be expected during the passing of the air across the foothills would be only one or two degrees (Lilly and Zipser 1972) which is almost within the range of uncertainty in the temperature data (± 1°C). The mean difference between the 'slope' and the 'Boulder' potential temperature (averaged over three hours) during the height of the storms was actually found to be -0.4°C which indicates some cooling, although considering the accuracy of the data the value should be rounded off to 0°C.

9. Surface Wind

No systematic study of surface winds during Boulder wind storms has ever been made. Ives, analyzing Chinook winds in this area and selecting his cases on the basis of temperature and precipitation criteria, found the average speed at Denver "during the warmest four hours of a Chinook" to be about 7-9 m s^{-1} and did not believe the tale of "anemometers that registered 75 mph (34 m s^{-1}) and then took off to Kansas like a helicopter" (Ives 1950, p. 321). It is now a known fact that strong wind storms, with gusts of over 45 m s^{-1} and causing property damages that can exceed $1 million, are not a unique occurence in Boulder (Julian and Julian 1969; Lilly and Zipser 1972). From casual observations it is also known that these wind storms are a local phenomenon and do not extend far into the plains to the east. Places like Denver (35 km to the east) are usually affected only by the most severe winds. It is also believed that the winds generally extend farther into the plains when the storm occurs during the day than at night because of additional energy required to remove a surface inversion; similar considerations would suggest a difference in the eastward extent between the warm and cold storms. Regarding wind speeds on the slopes to the west, the popular assumption is that there is a close relationship, i.e., during wind storms in Boulder the wind on the slopes is believed to be strong also.

As a first step toward the study of the distribution of surface wind speeds during storms in Boulder, computer plots of the 5-minute wind data (mean, minimum and gust speeds) are presented for the 20 wind storm cases in Figs. 9.1 to 9.20. These plots are excellent reproductions, in condensed form, of the actual wind traces. For each storm the traces from the east-west line of 7 anemometer stations, centered on Boulder, are arranged with respect to location such that the one on top comes from the highest mountain station shown in the diagram which is also the one furthest west of Boulder and the bottom trace represents the lowest station which is also furthest to the east. It is important to note, however, that although the traces are equally spaced in the diagram, the actual distances between the stations are not equal, this is particularly true with respect to the center traces (Jones, Southern Hills and 30th Street) which are all in Boulder.

It is evident from these plots that there is a great diversity in wind storm pattern. Some lasted only a few hours while others

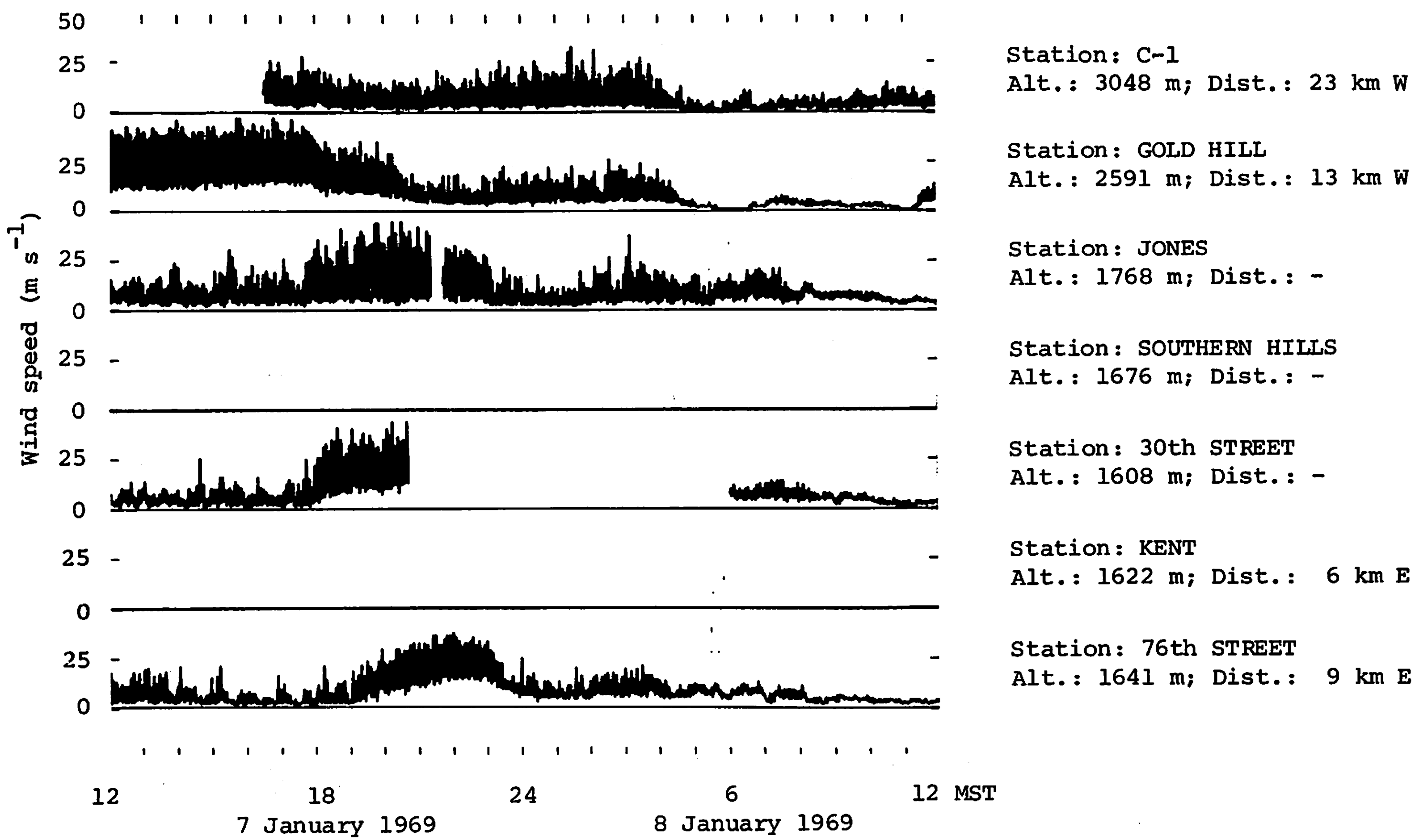


Fig. 9.1. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 7 January 1969 wind storm.

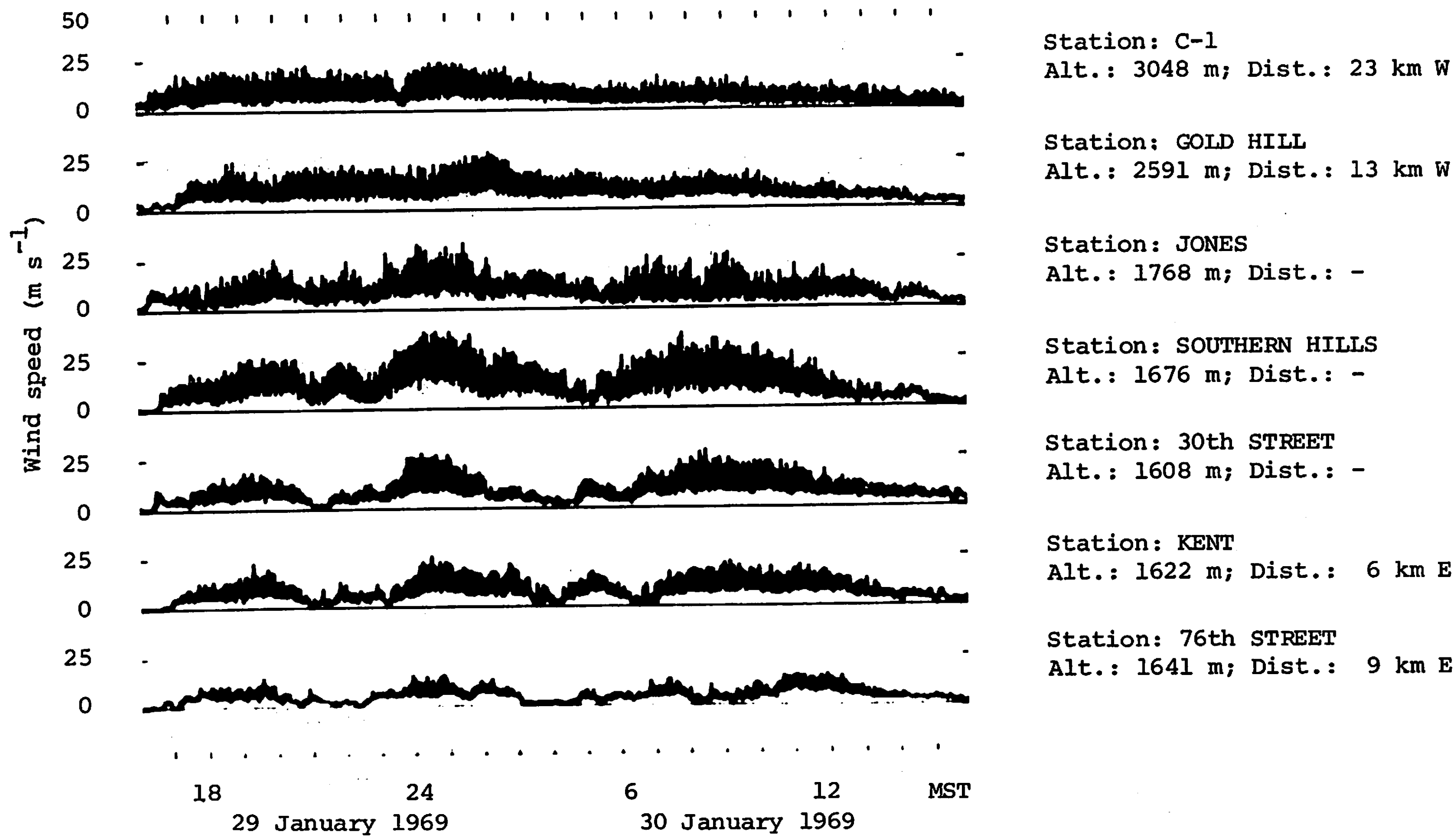


Fig. 9.2. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 29-30 January 1969 wind storm.

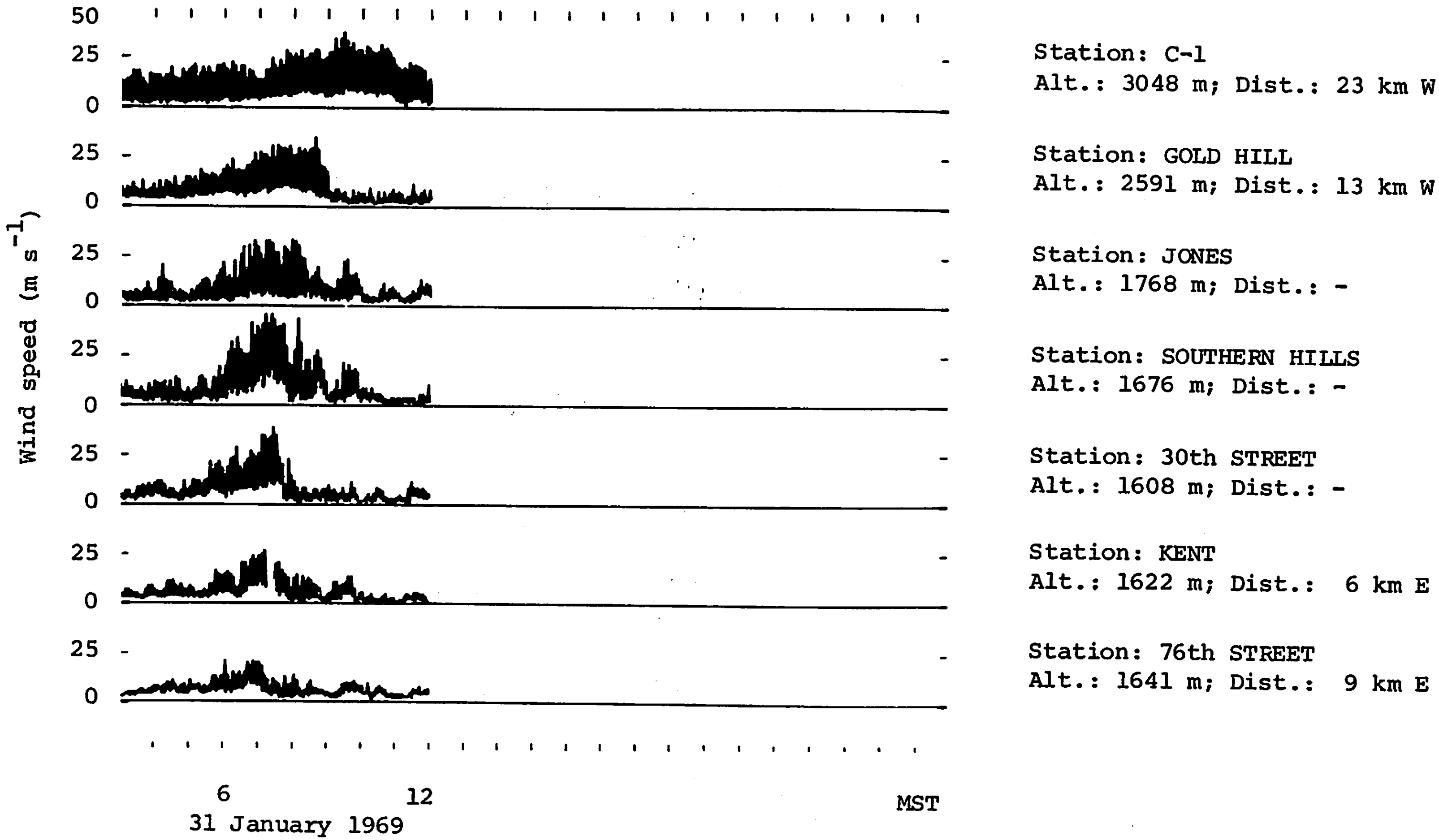


Fig. 9.3. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 31 January 1969 wind storm.

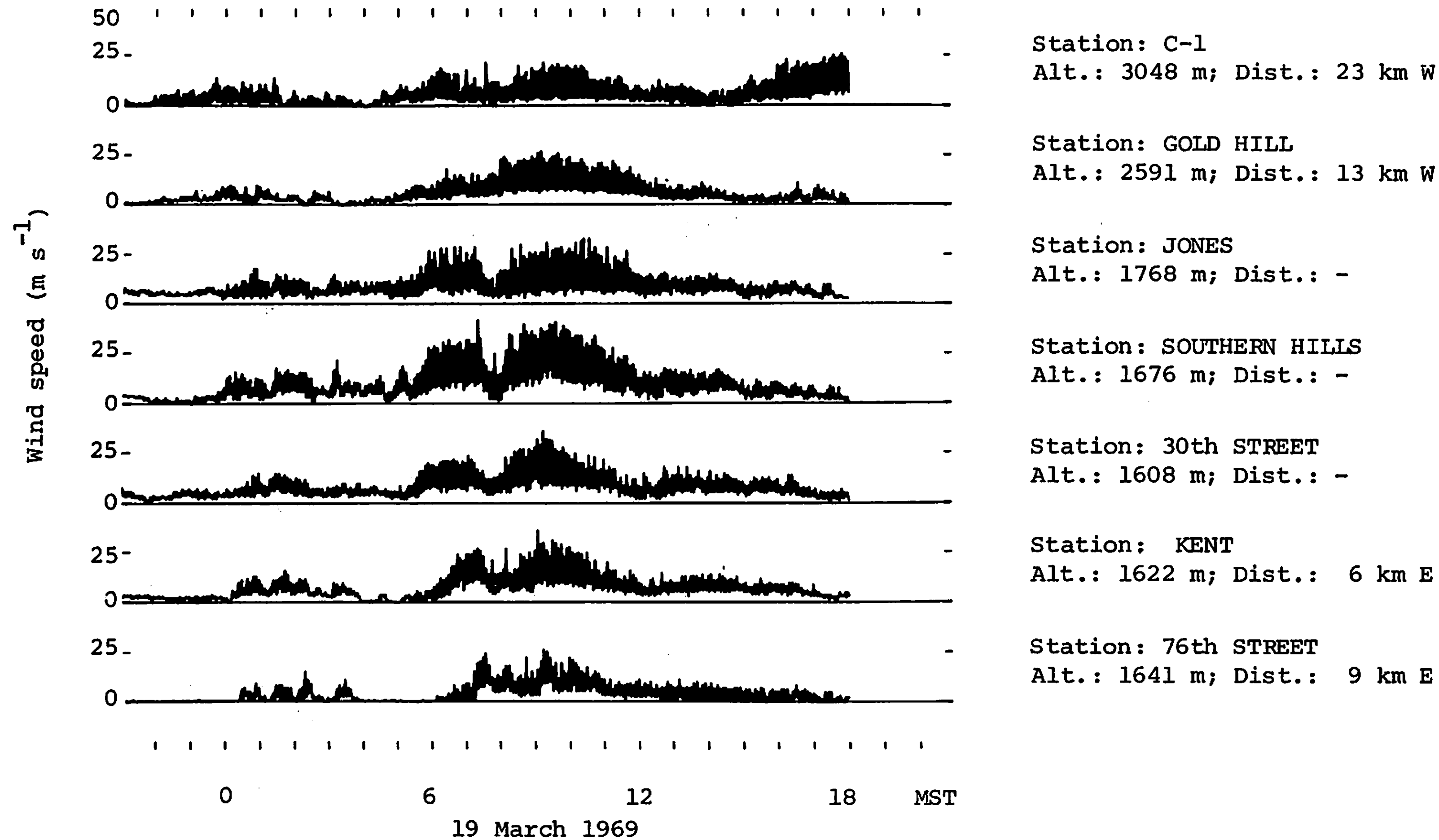


Fig. 9.4. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 19 March 1969 wind storm.

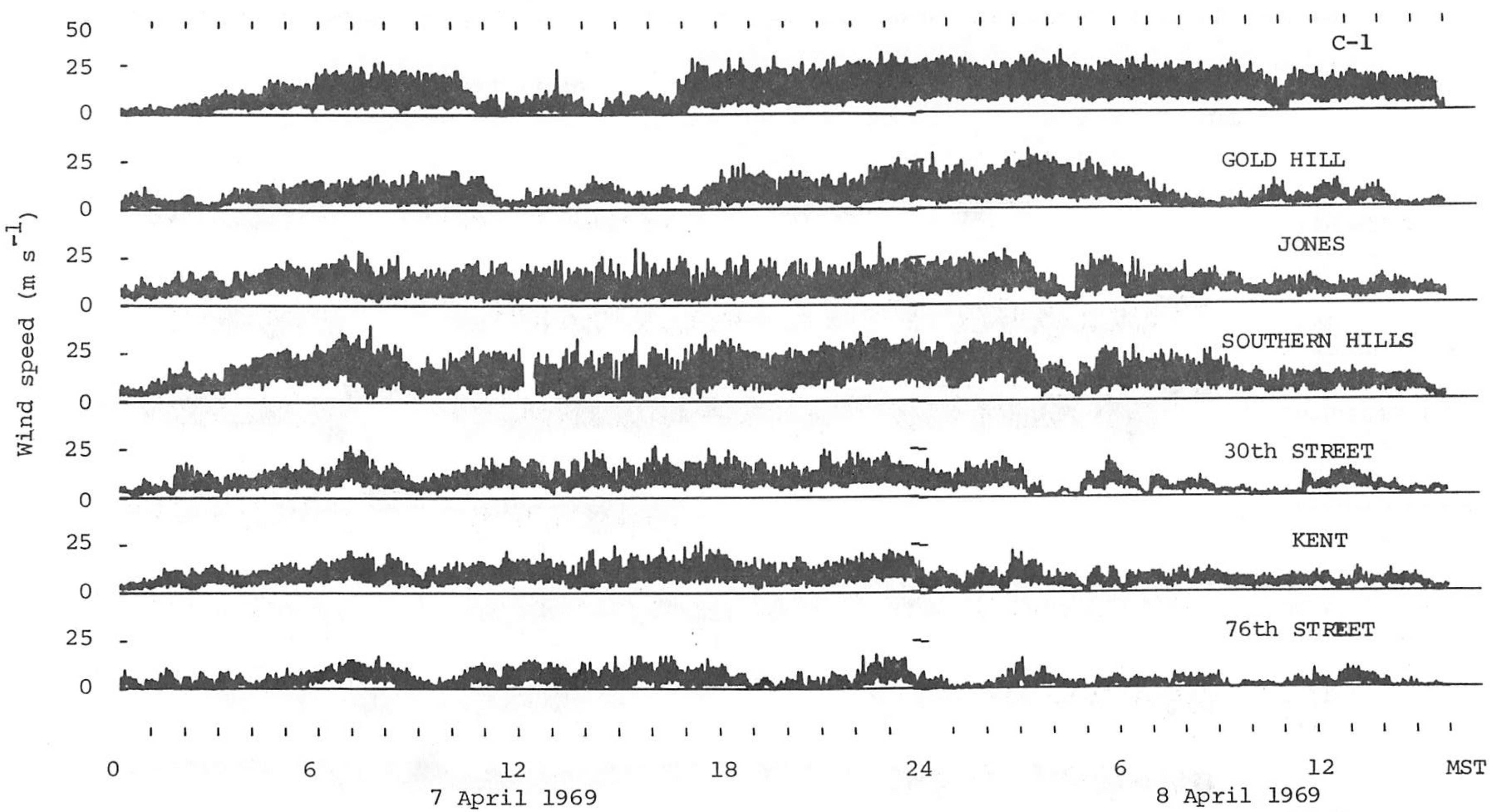


Fig. 9.5. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 7-8 April 1969 wind storm.

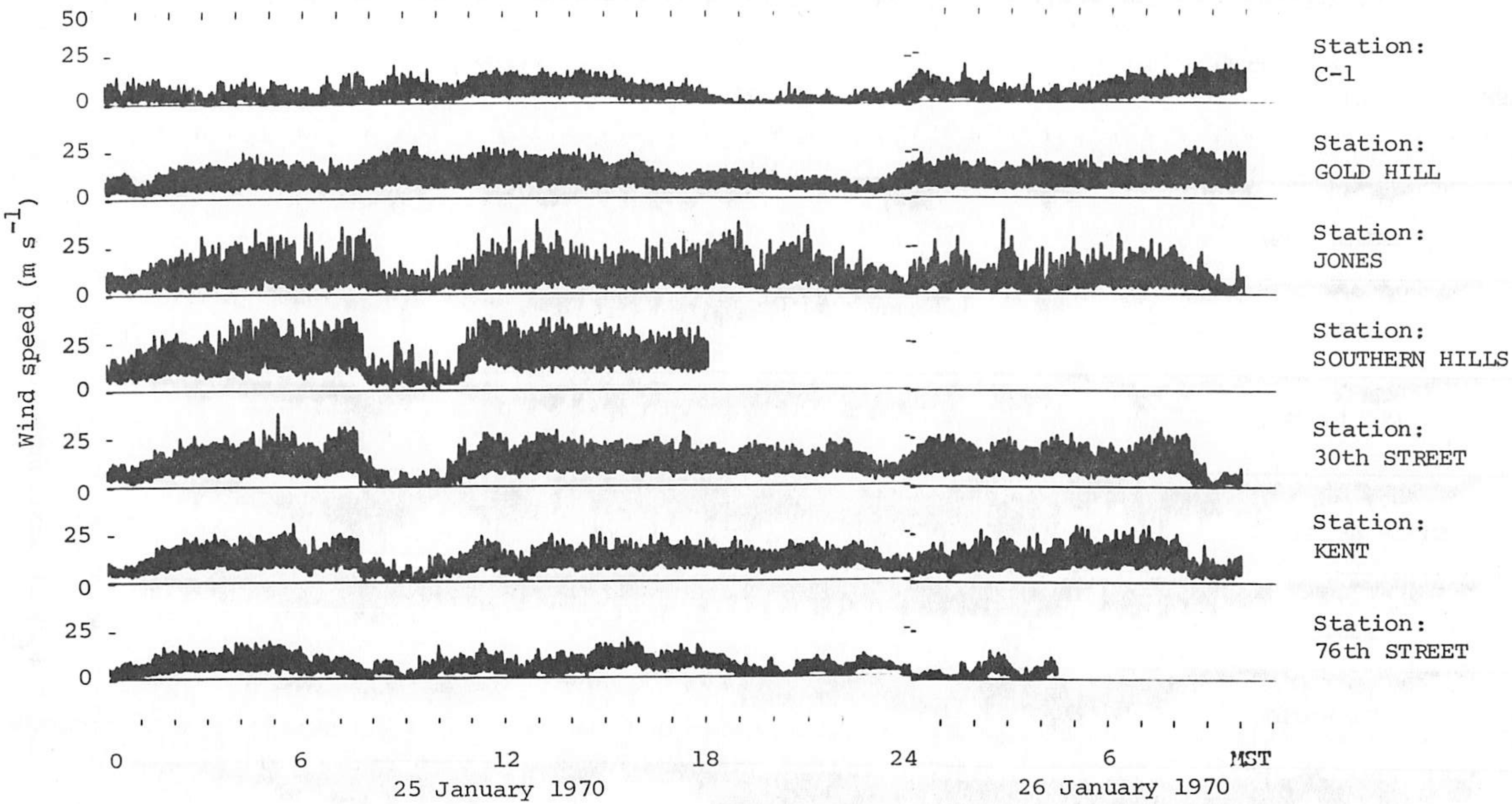


Fig. 9.6. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 25-26 January 1970 wind storm.

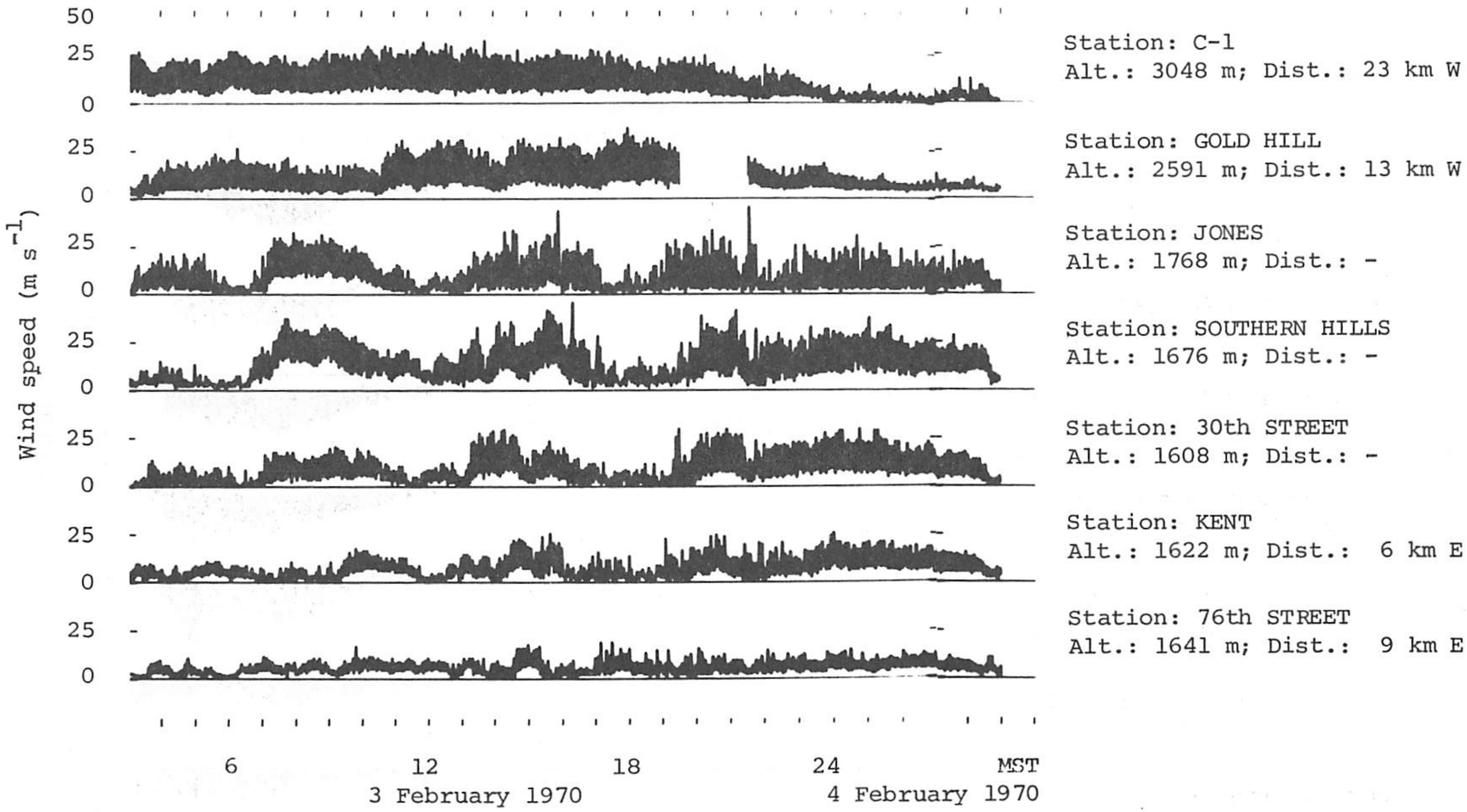


Fig. 9.7. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 3-4 February 1970 wind storm.

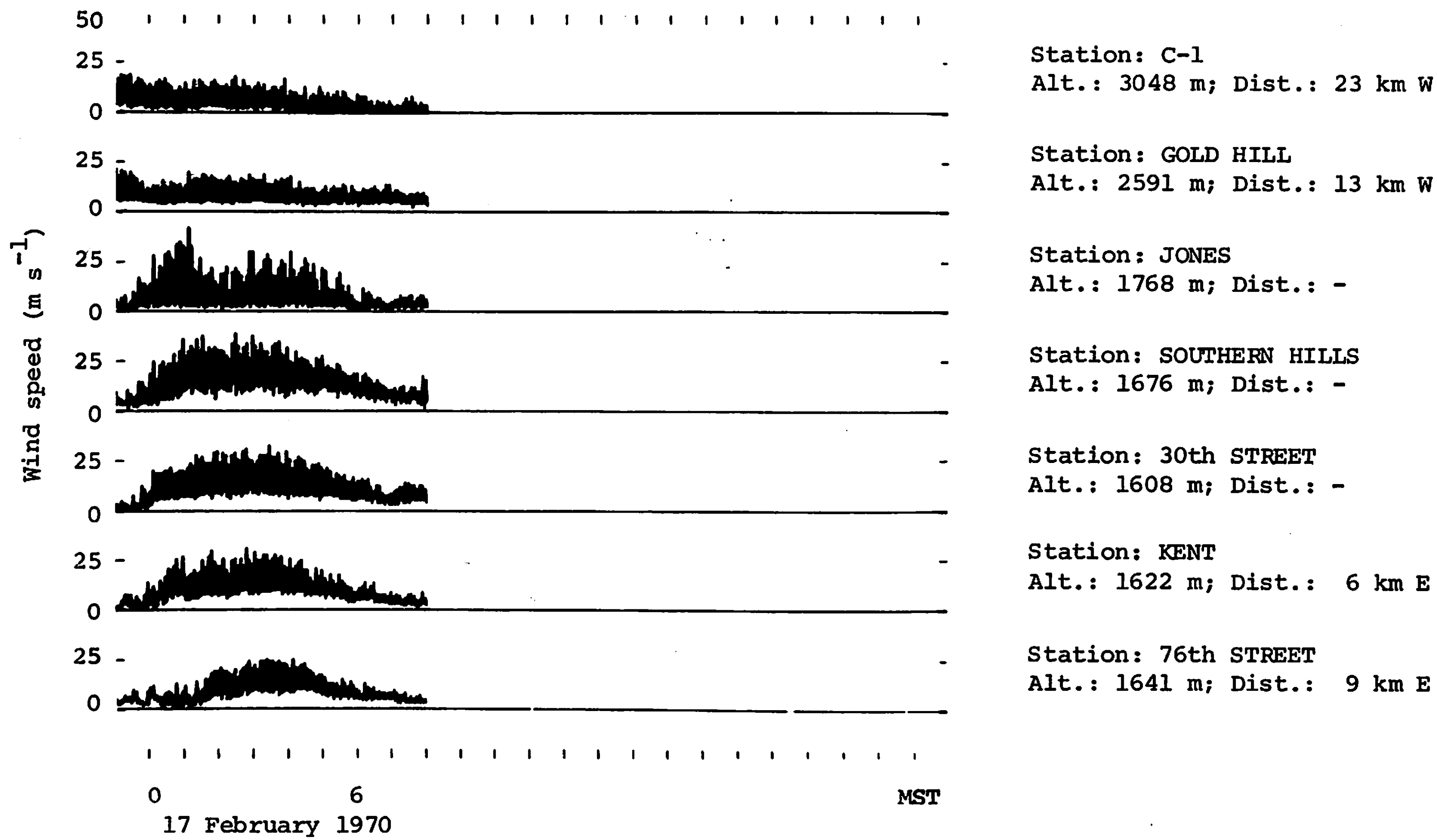


Fig. 9.8. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 17 February 1970 wind storm.

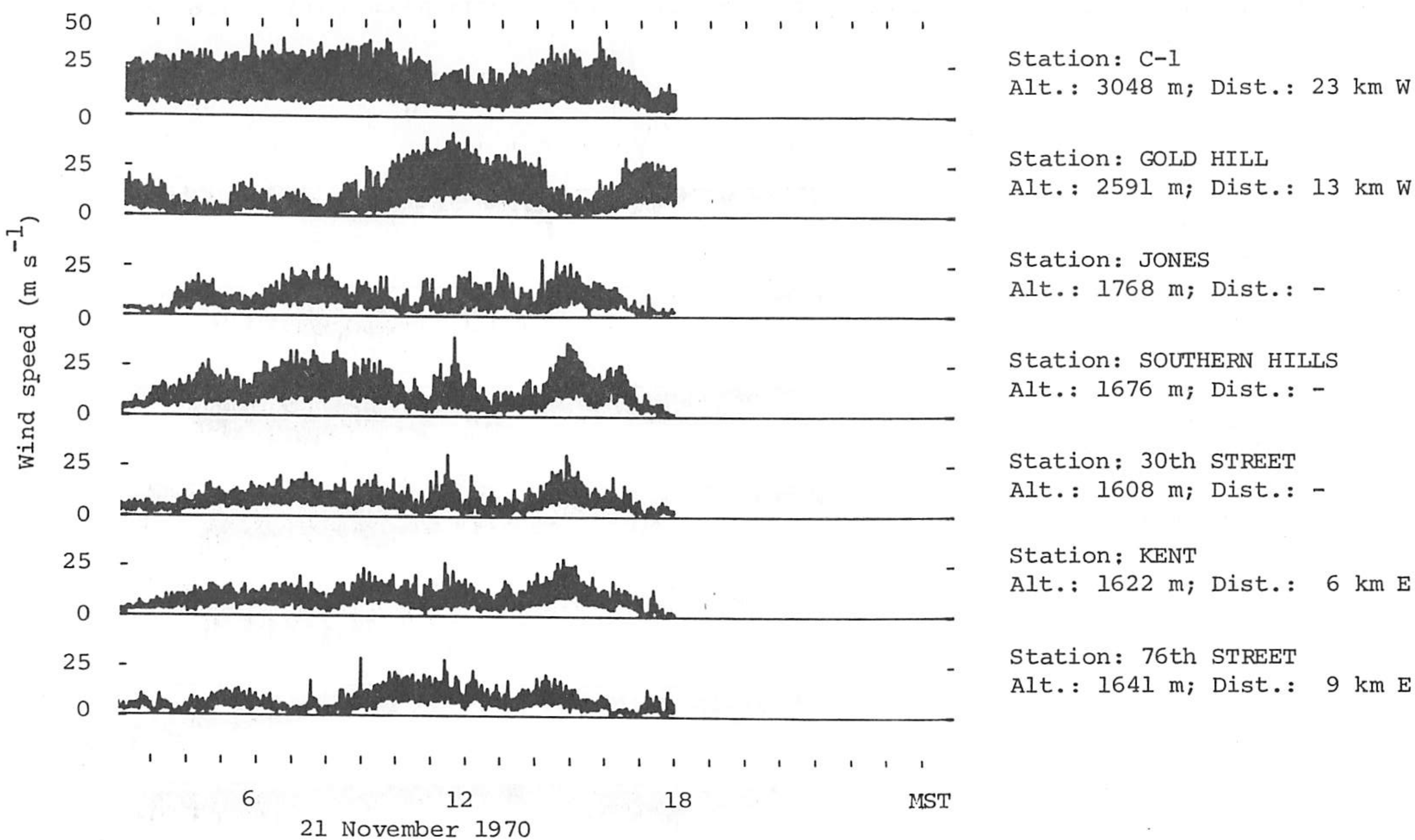


Fig. 9.9. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 21 November 1970 wind storm.

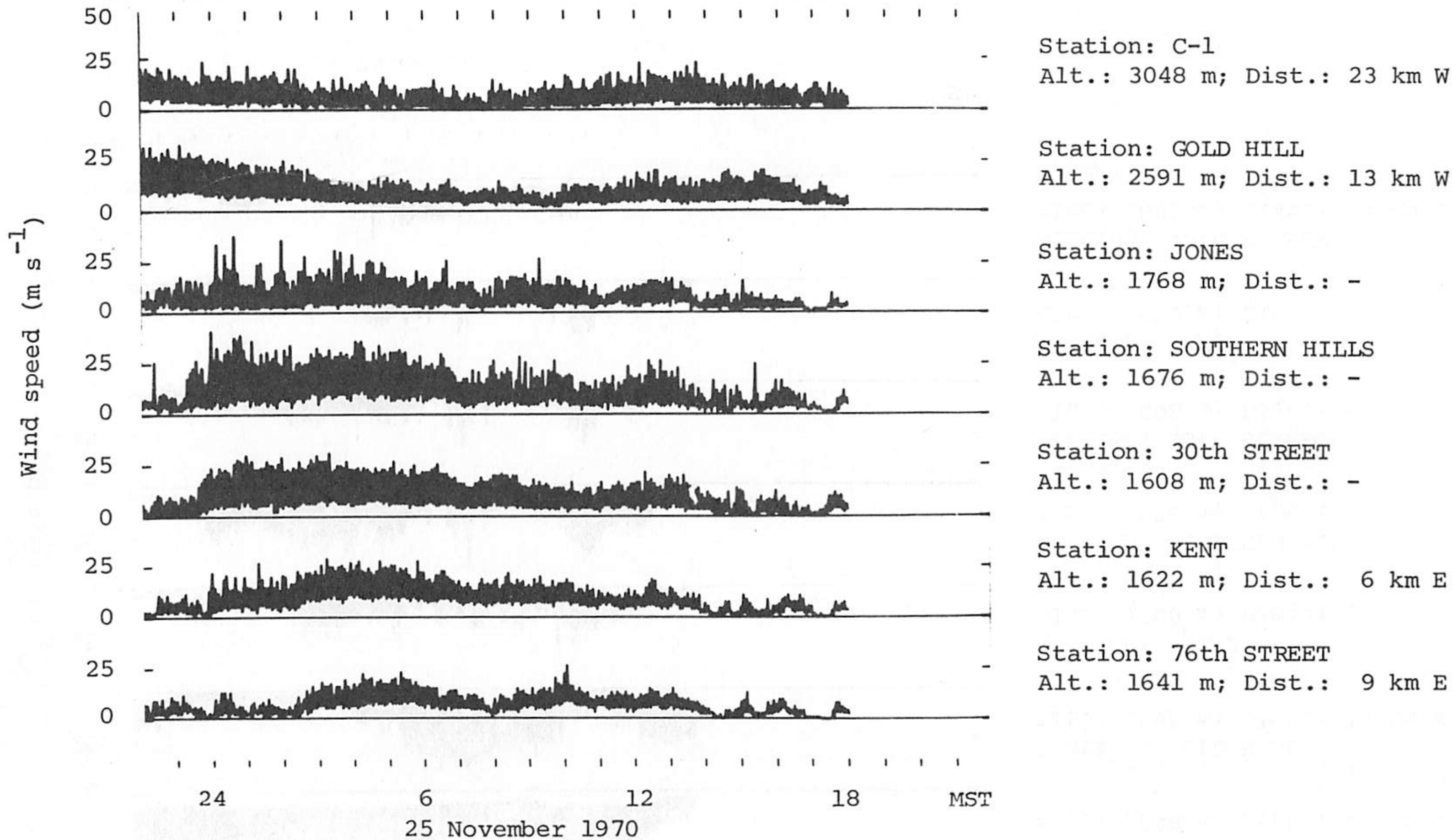


Fig. 9.10. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 25 November 1970 wind storm.

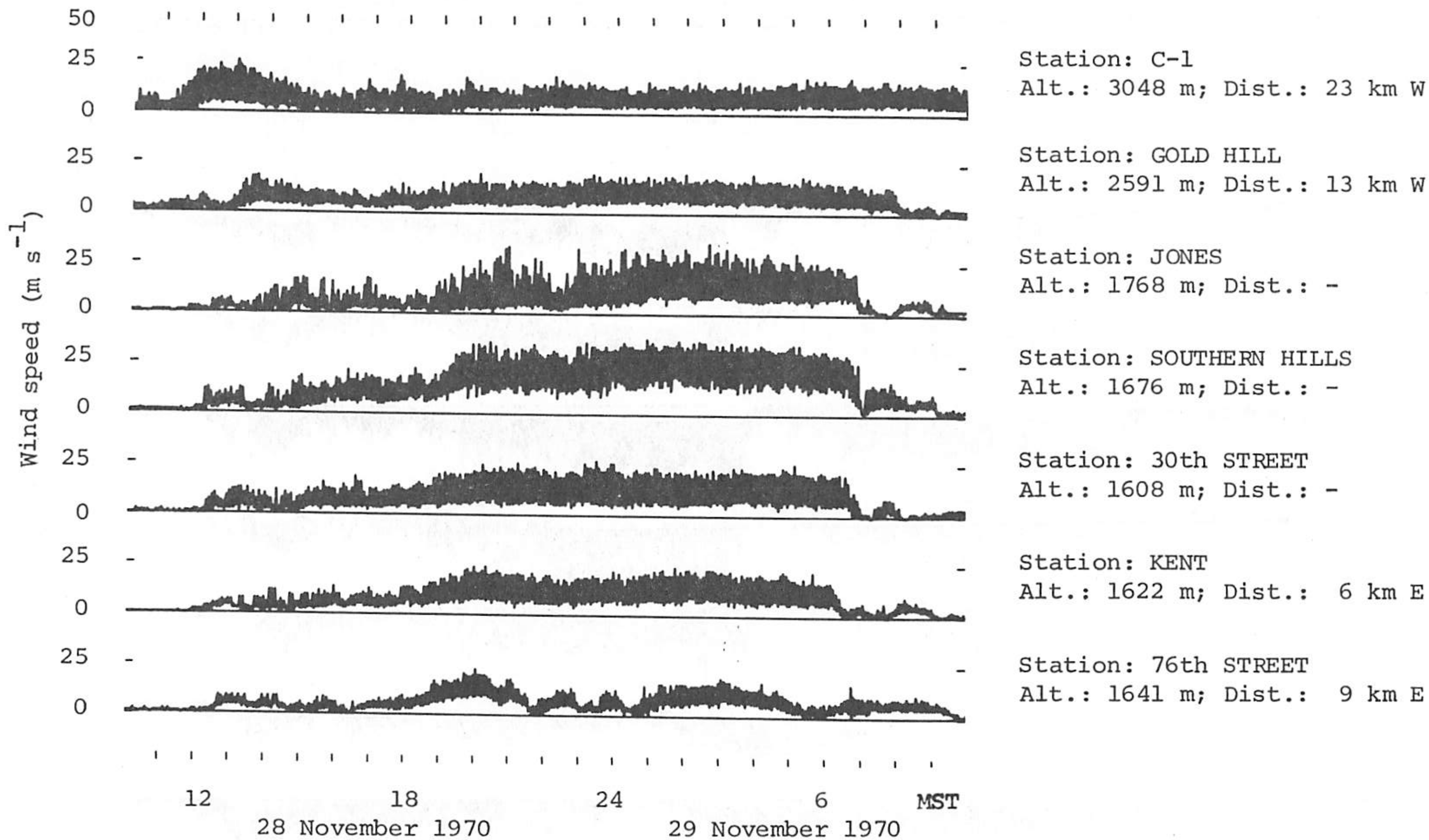


Fig. 9.11. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 28-29 November 1970 wind storm.

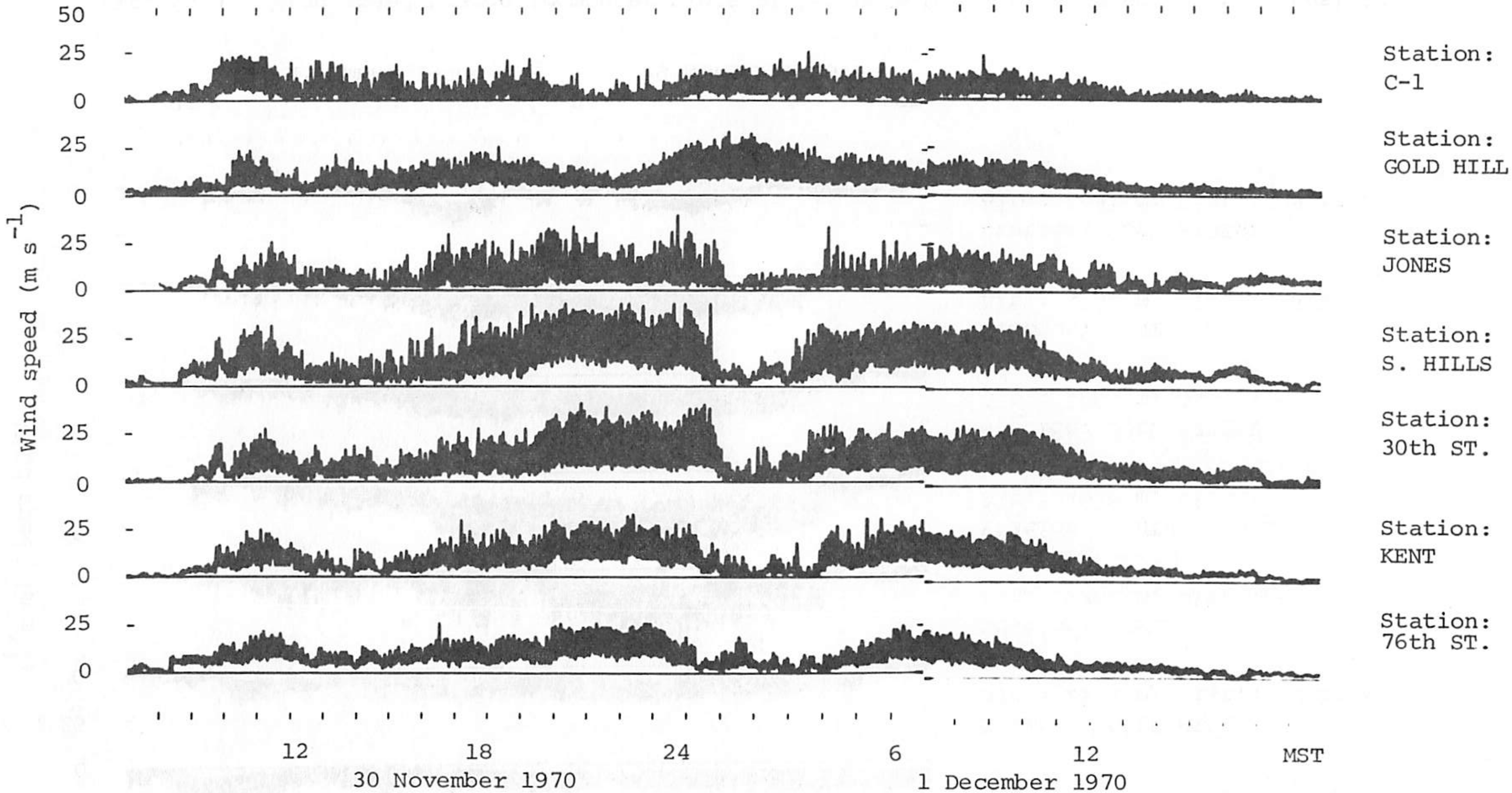


Fig. 9.12. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 30 November-1 December 1970 wind storm.

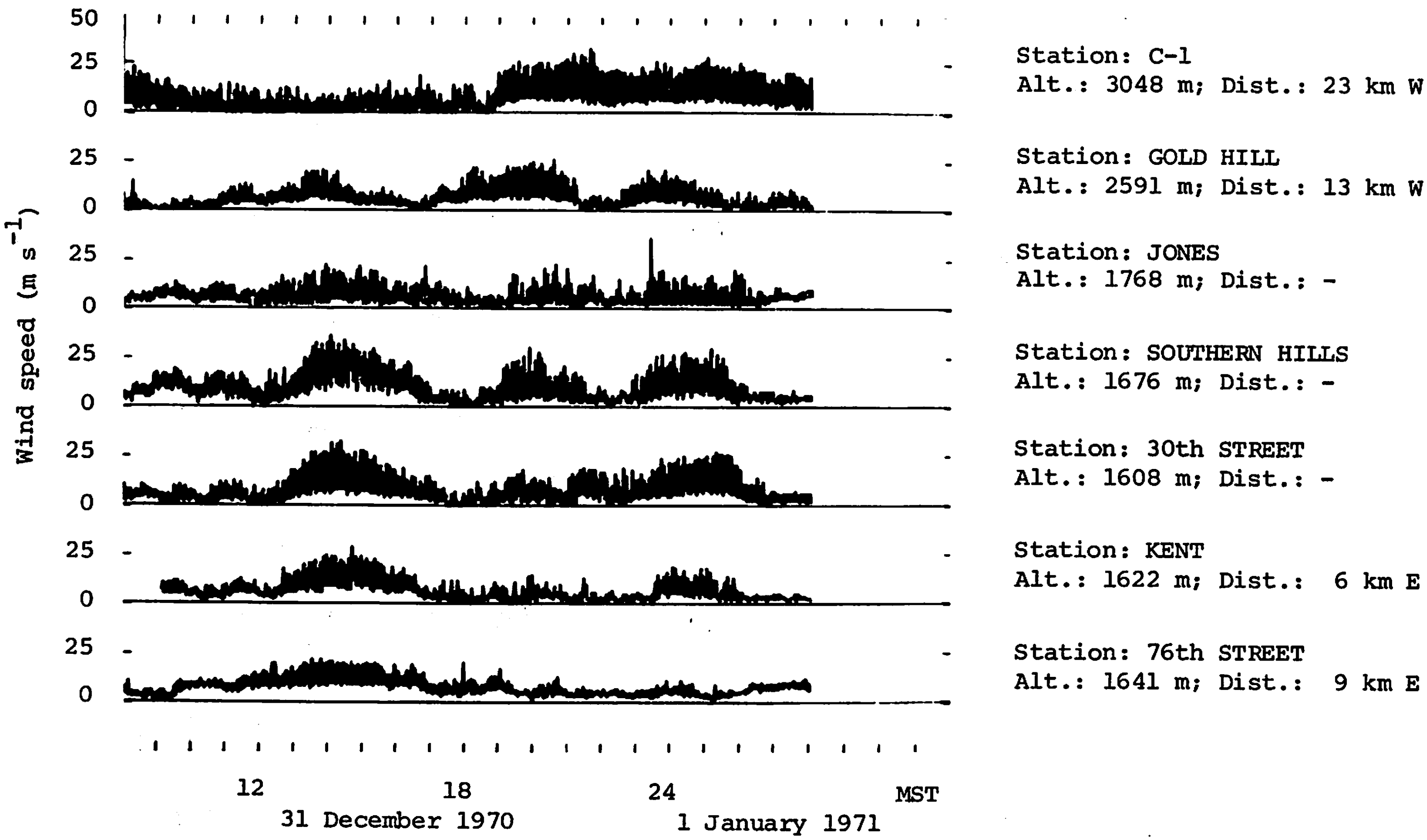


Fig. 9.13. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 31 December 1970 wind storm.

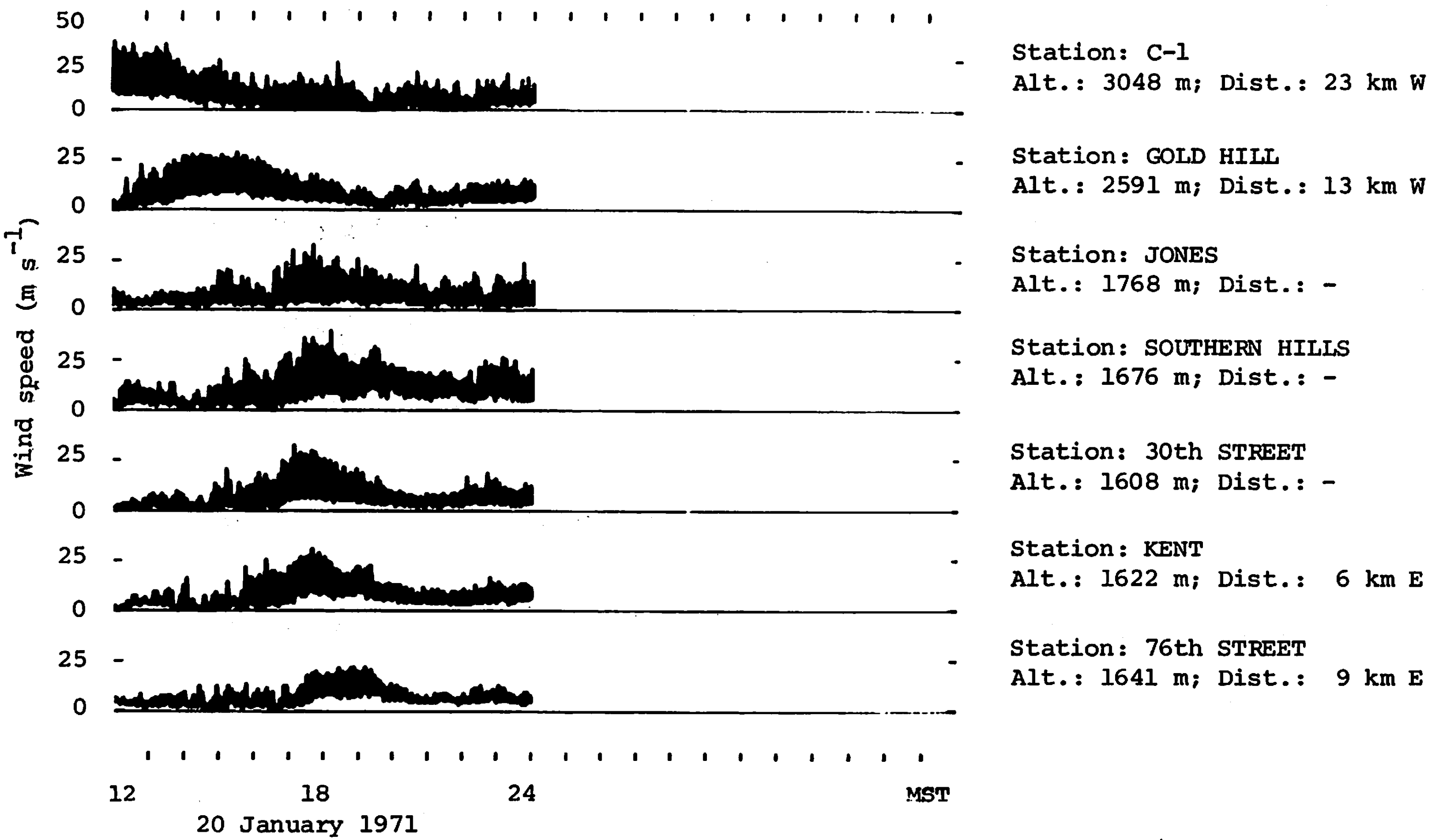


Fig. 9.14. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 20 January 1971 wind storm.

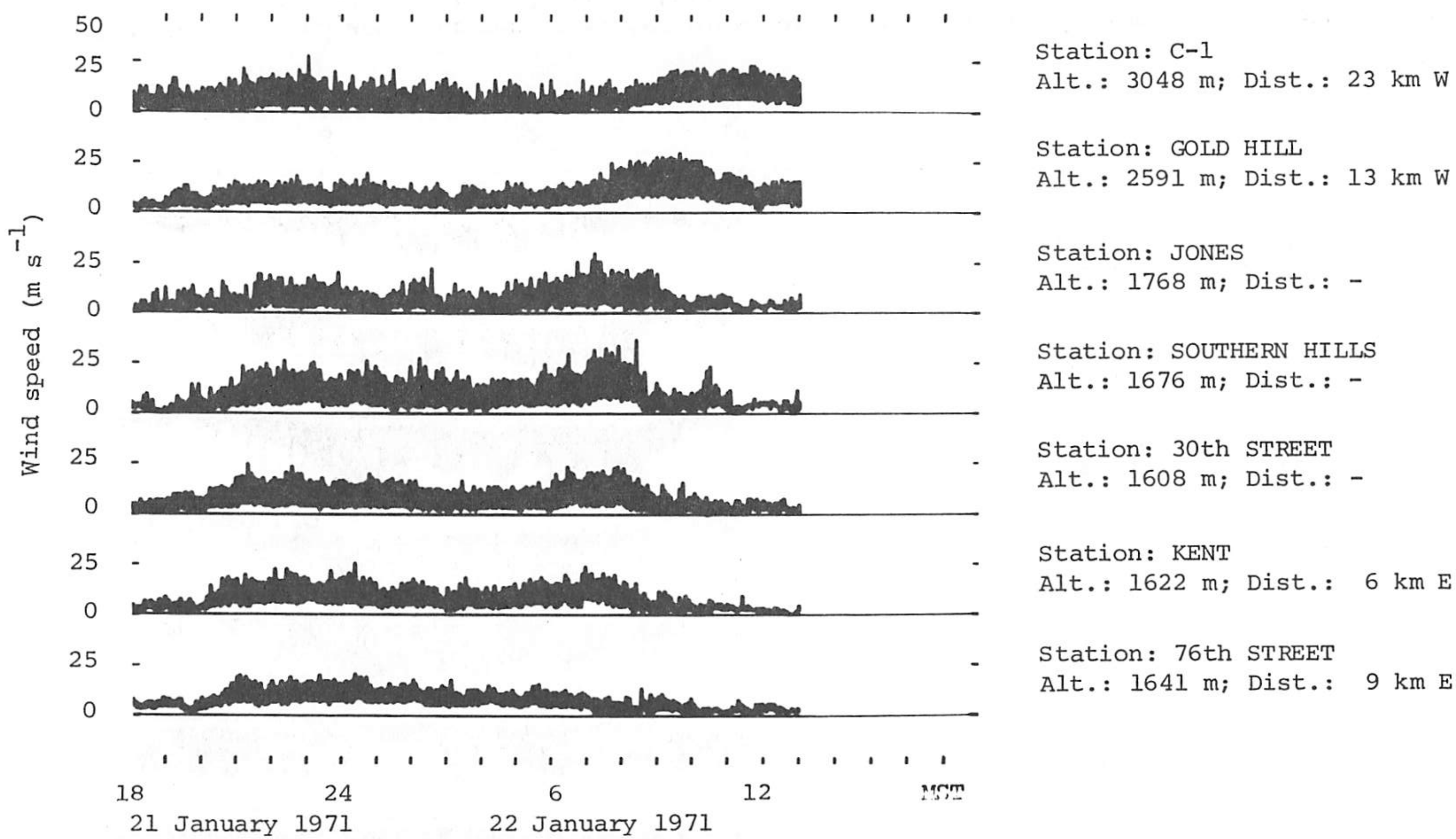


Fig. 9.15. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 22 January 1971 wind storm.

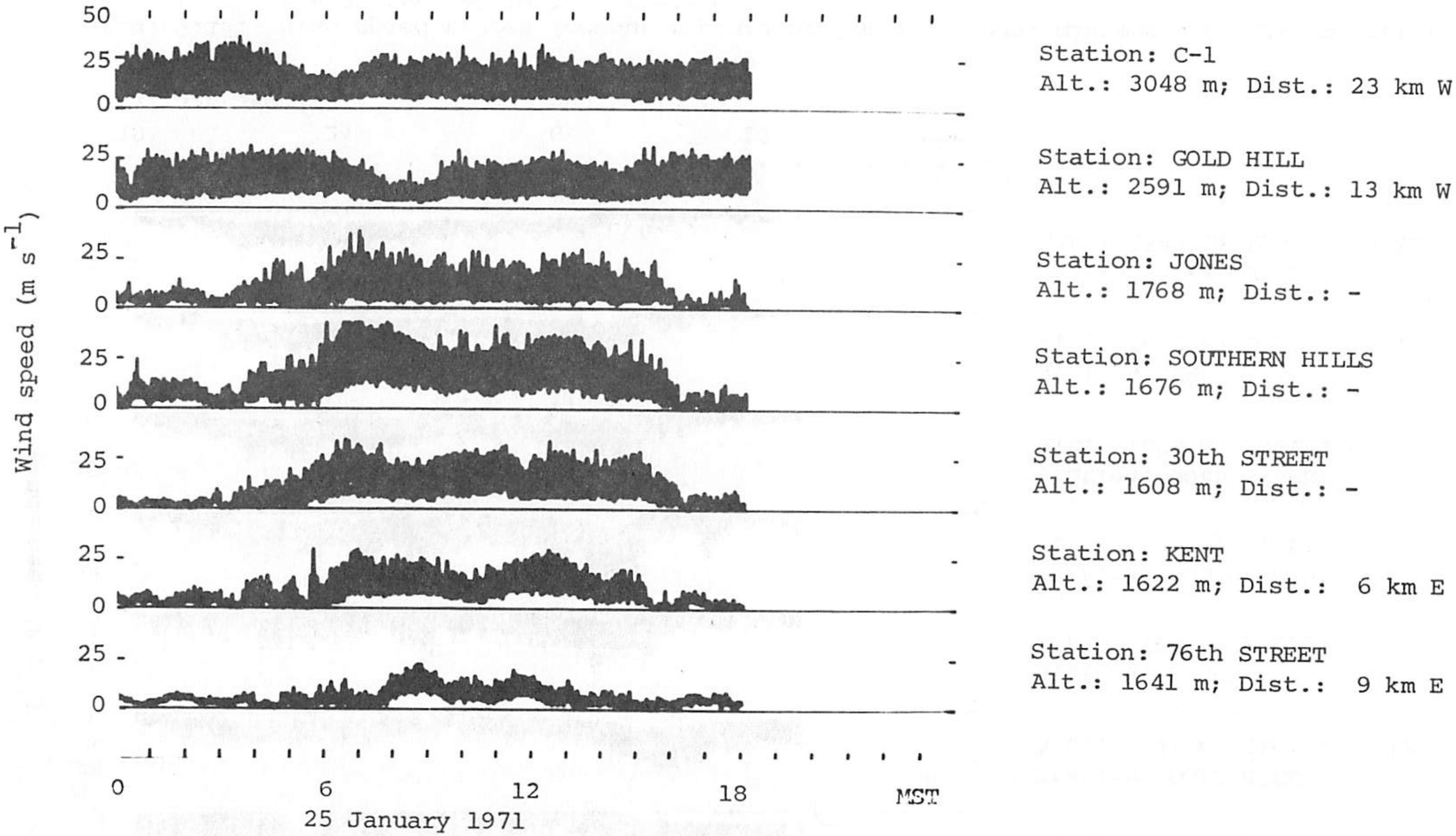


Fig. 9.16. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 25 January 1971 wind storm.

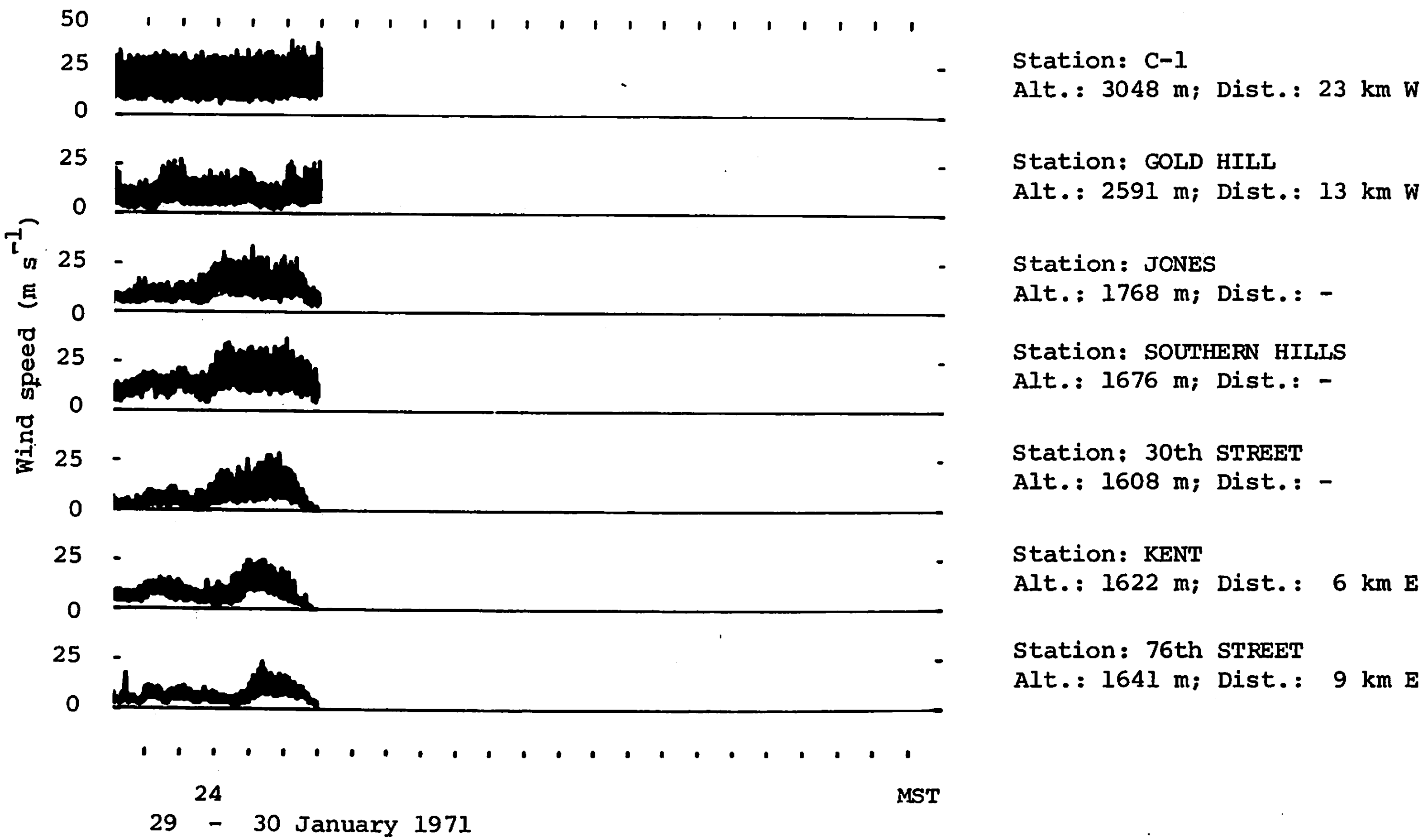


Fig. 9.17. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 29-30 January 1971 wind storm.

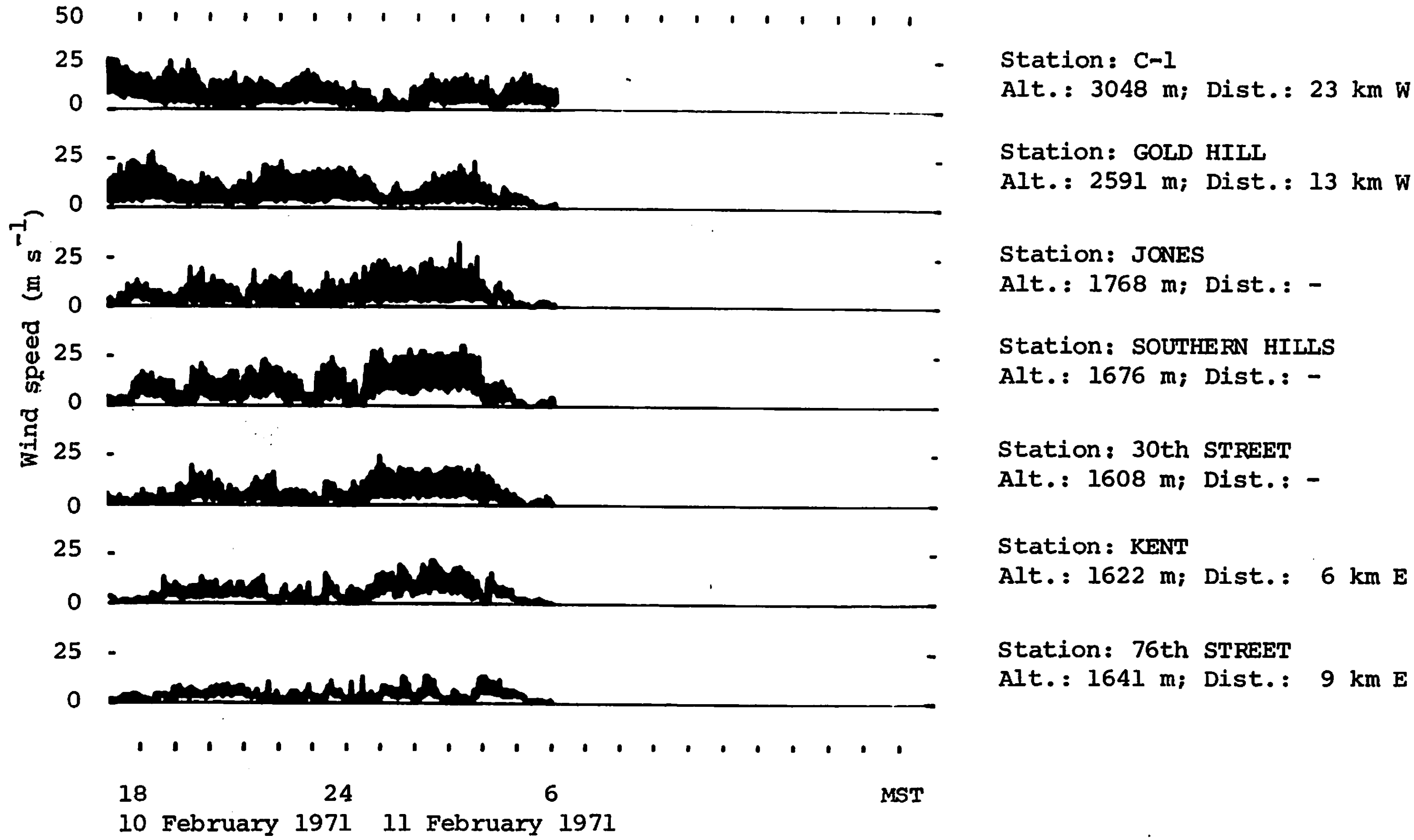


Fig. 9.18. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 10-11 February 1971 wind storm.

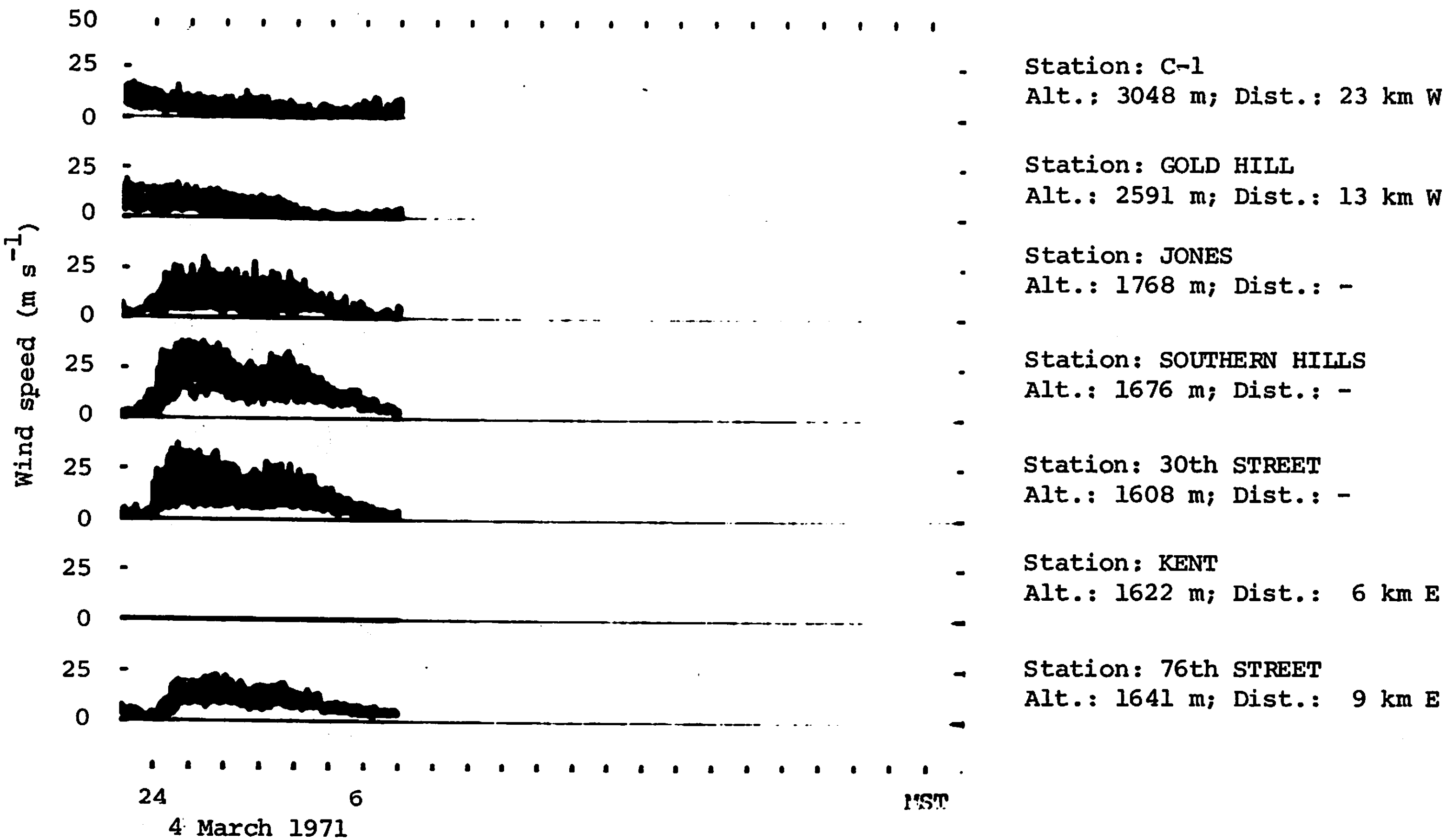


Fig. 9.19. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 4 March 1971 wind storm.

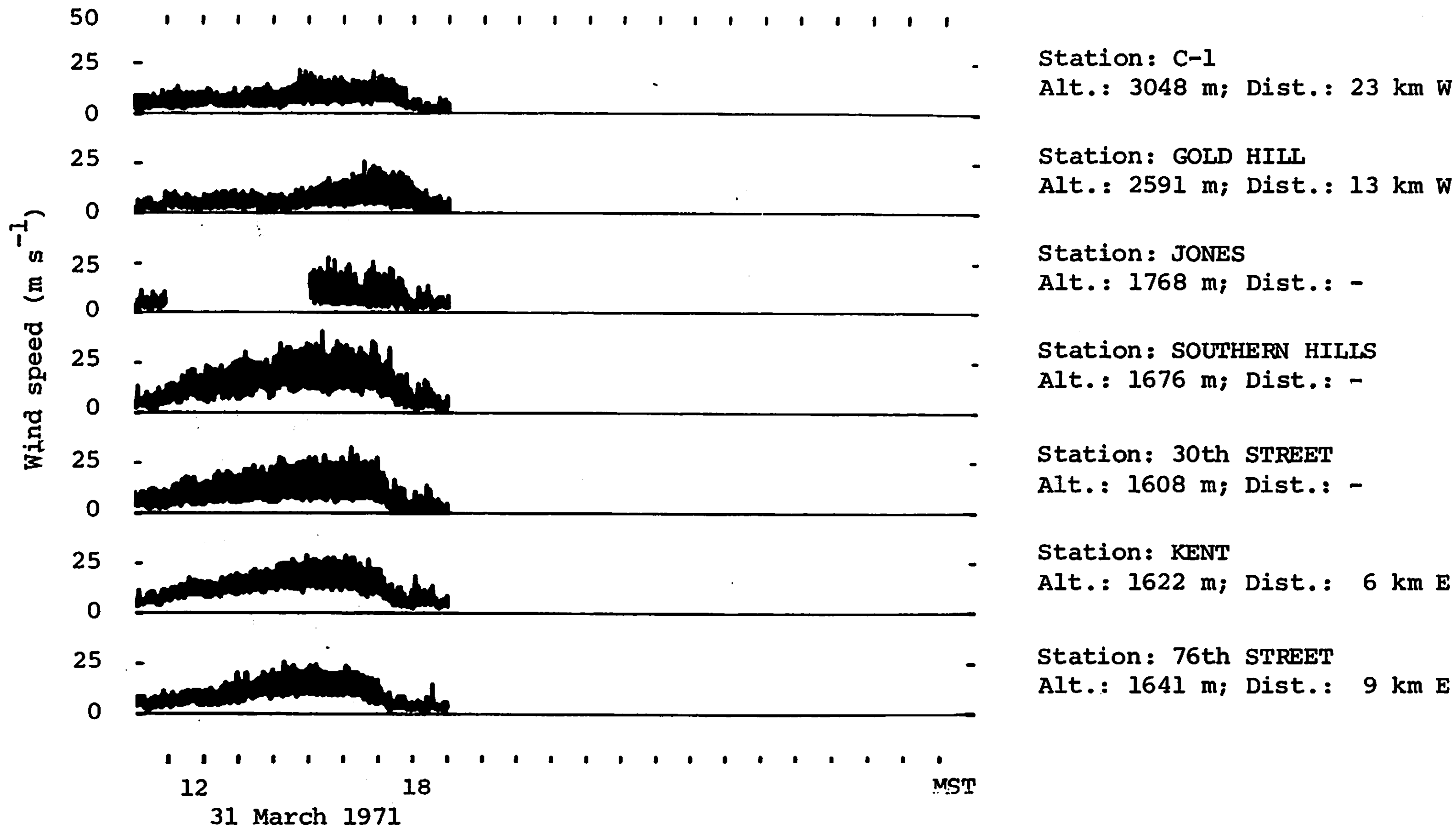


Fig. 9.20. Wind speed traces (computer plots of 5-minute mean, minimum and gust speeds) for the 31 March 1971 wind storm.

continued for more than a day (for example, the 31 January 1969 storm vs. that of 7-8 April 1969). In some cases the wind was strong at the Boulder stations throughout the entire storm period while in others interruptions or 'pauses', in terms of marked periods of relatively low velocities occurred, lasting for several hours (for example, the 28-29 November 1970 storm vs. that of 25 January 1970). The relationship between the 7 stations appears to be rather complex too. Although the wind speed pattern at the three Boulder stations (Jones, Southern Hills and 30th Street) is generally similar, with more or less simultaneous onsets and cessations of the storms and pauses, this changes with distance to the east. Not only is there an increasing time lag between the Boulder stations, Kent (about 6 km to the east) and 76th Street (about 9 km to the east), but the whole pattern of increases and decreases in speed can occasionally take on a different form. In the case of the 3-4 February 1970 storm, for instance, there is hardly any indication at the 76th Street station of the marked changes in wind speed taking place in Boulder; at the former the velocity was relatively low throughout the entire wind storm period in Boulder. In the 28-29 November 1970 case, on the other hand, marked wind speed changes are noticeable at the 76th Street station while in Boulder and at Kent the wind was remarkably steady throughout the storm. Differences between wind patterns of the Boulder and east slope stations are even more striking. In some cases the wind storms and pauses in Boulder do not seem to be at all reflected in the wind speed traces at the slope stations (examples of this are the 17 February 1970 and 25 January 1971 storms), while in others the relationship seems to be an inverse one, suggesting an apparent 'movement of a wind speed maximum'. A good example of this movement is the 20 January 1971 wind storm. As the wind speed at C-1 (the uppermost slope station in the diagram, about 23 km west of and 1450 m above Boulder) decreased, it increased at Gold Hill (about 13 km west of and 980 m above Boulder); and as the speed decreased again at Gold Hill the wind storm began in Boulder. This kind of movement of a wind speed maximum is different from a mere time lag in the onset and cessation, which would still result in a simultaneous maximum at all stations.

A closer examination of the wind speed plots suggests that at least some of the diversity in wind speed patterns can be explained in terms of such 'movements' of wind maxima. Fig. 9.21 shows schematic wind speed traces from three neighbouring stations in an east-west line parallel to the direction of the general air flow. The stations are assumed to be arranged in a manner similar to that of the actual 5-minute wind data plots discussed above, with the top trace representing the highest mountain station which is also the station farthest to the west. In the diagram, Type I represents the wind speed pattern resulting from a downslope or eastward moving maximum, Type II an upslope or westward moving one. Examples of Type I can be seen in the traces for the 7 January 1969 and 20 January 1971 wind storms and of Type II in the 31 January 1969 and 22 January 1971 storms. Combination Type I/II results in a pause at the upper or westward station, a merging of the peaks and decreasing duration of this peak at the lowest station, farthest to the east. Some indication of this can be seen in the wind speed

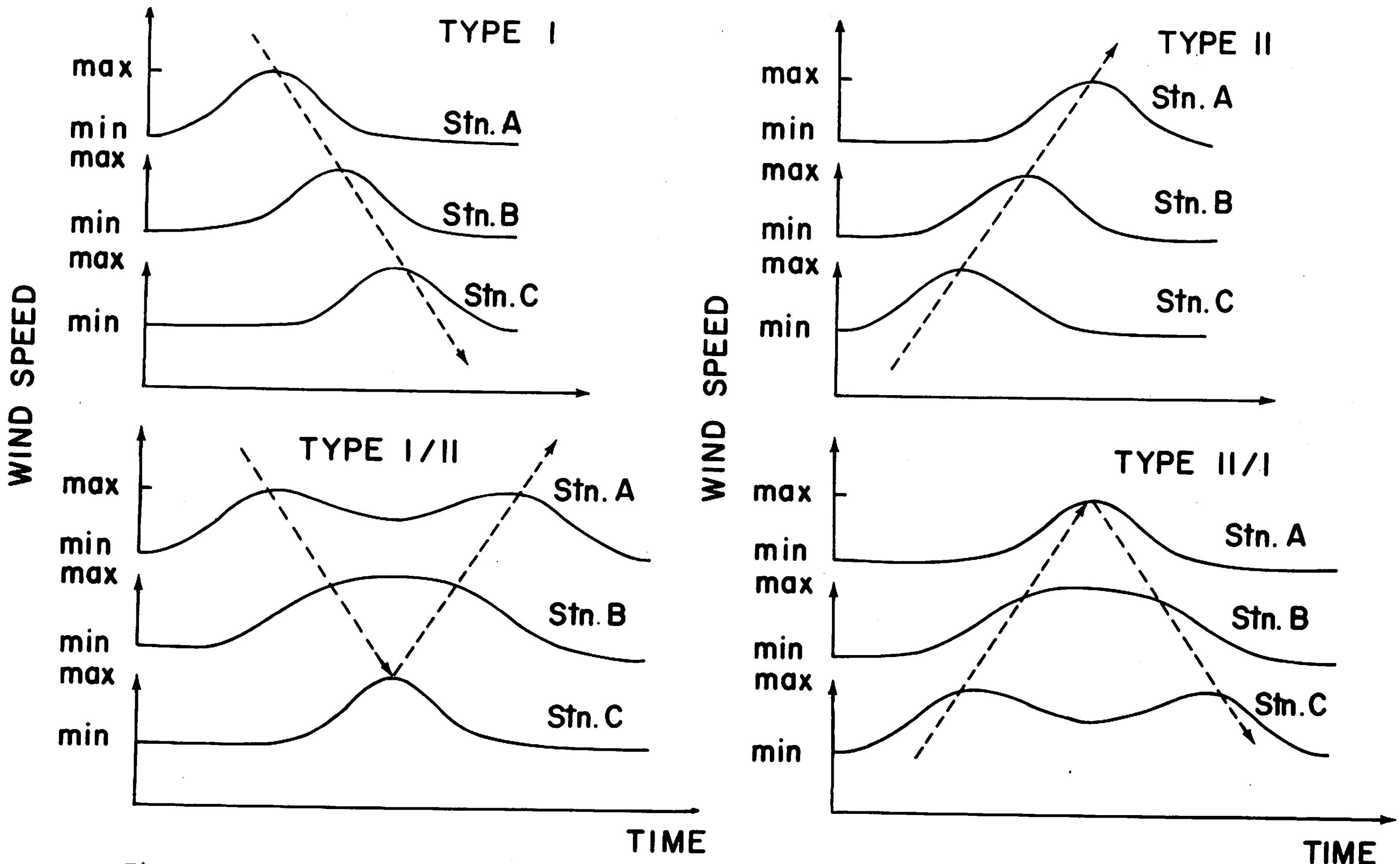


Fig. 9.21. Schematic diagrams of the effect of a moving wind maximum on the mean wind speed traces of three neighbouring stations.

plots for the 25 November 1970 wind storm. Combination Type II/I, the reverse of Type I/II with a pause at the lowest and most eastward station, apparently occurred during the 30 November-1 December 1970 wind storm. The clarity of the pattern depends of course on the speed of the phenomenon, the slower the movement the more outstanding the pattern. Furthermore, a station situated near the edge of the wind maximum will show larger variations than the others, in response to slight changes in the position of the maximum. An example of the latter is the 28-29 November 1970 case during which the eastern edge of the wind maximum appears to have been situated near the 76th Street station. The 21 November 1970 case seems to be an example of the simultaneous occurrence of two maxima: wind speeds at the Boulder stations were out-of-phase with Gold Hill and to some degree also with 76th Street, while the wind maximum at Gold Hill clearly coincided with a minimum at C-1. The 19 March 1969 storm, on the other hand, is an example of the almost simultaneous occurrence of the wind at all stations which is perhaps noticeable only because of the very low initial speeds at the mountain stations since the actual increase in wind speed at these stations during the storm in Boulder was relatively small, certainly much less than in Boulder where the wind exceeded hurricane force.

The cause of the moving wind speed maxima is demonstrated in Fig. 9.22 which shows two potential temperature fields (representative of streamlines in dry-adiabatic flow) as observed during the 22 January 1970 wind storm by two NCAR airplanes flying back and forth across the Continental Divide (the author being one

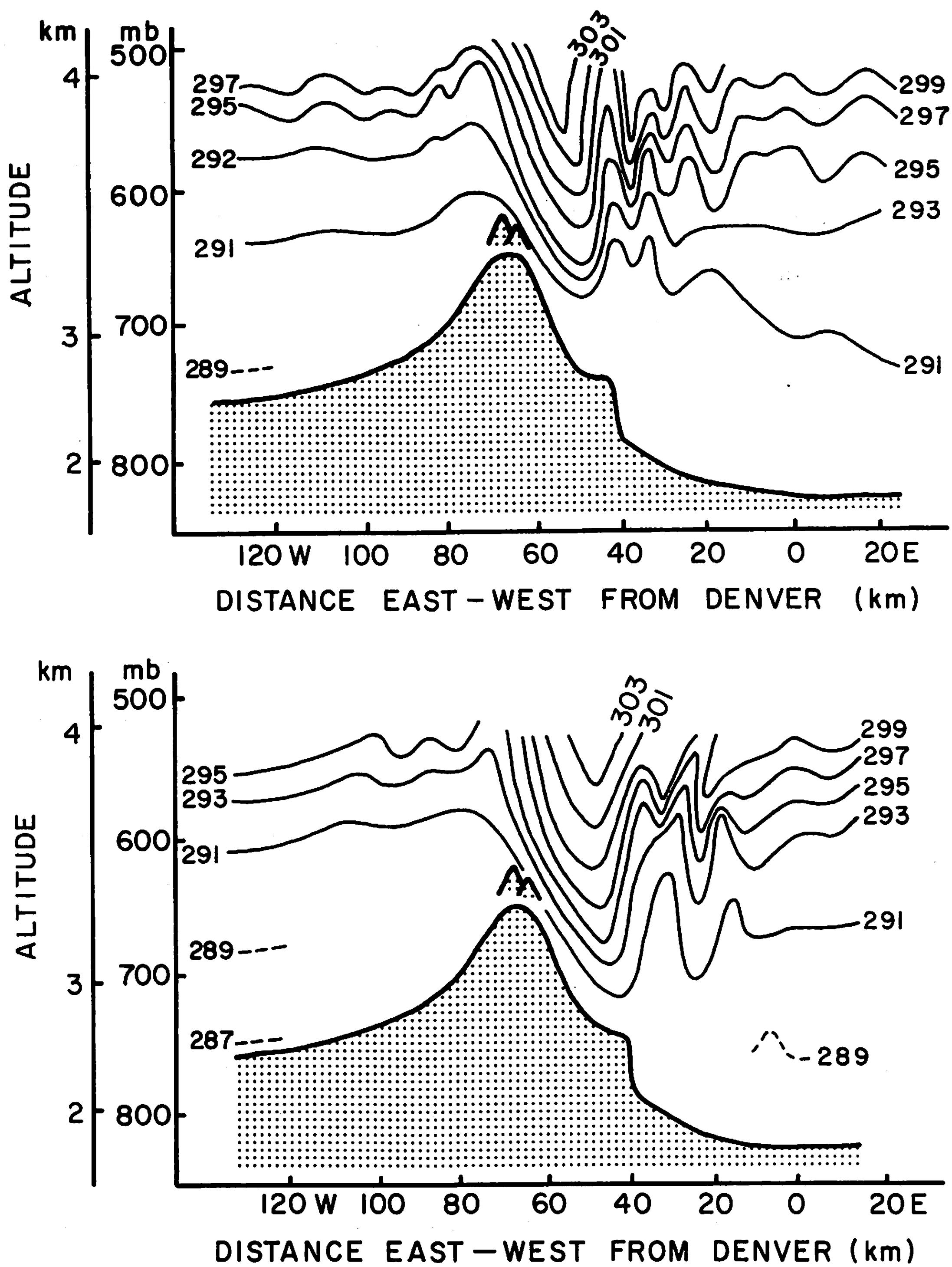


Fig. 9.22. Cross section of the potential temperature field in degrees K on 22 January 1970. Bottom: 10:00-11:00 MST, top: 12:00-13:00 MST.

of the observers). The recorded potential temperature vs. static pressure was obtained by a special computer program (Lilly et al. 1971) and the results were plotted along the flight tracks. In the bottom cross section of Fig. 9.22, representing the potential temperature field between 10:00 and 11:00 MST, a 'lee air flow disturbance' is located over the foothills, slightly west of Boulder. Surface winds in Boulder had just decreased at that time but were at their maximum at the lowest mountain station, Gold Hill (Fig. 9.15). The top cross section in Fig. 9.22, representing the potential temperature field between 12:00 and 13:00 MST, shows the 'lee disturbance' moved westward and up the slopes. Surface winds in Boulder at that time were low, they were decreasing at Gold Hill but attaining their maximum at the next higher mountain station, C-1. Thus, the moving wind speed maximum, a Type II pattern as shown in Fig. 9.21, was a manifestation, at the surface, of the moving 'lee air flow disturbance' toward which winds are accelerated (as explained in Section 12).

The actual wind speed reached at the stations is another factor contributing to the diversity in wind patterns. During a given wind maximum in the Boulder area, speeds are generally highest at the Southern Hills station because of its open exposure (as compared with the other two Boulder stations) and decrease with increasing distance east. A good example of this can be seen in the wind traces for the 29-30 January 1969 storm. The relationship between wind speed in Boulder and at the two mountain stations, on the other hand, is less obvious. During a given storm in Boulder, wind speeds at the slope stations may be relatively low (as in the case of the 4 March 1971 wind storm) or high (for example, the 3-4 February 1970 storm). Comparison of the velocities reached during the actual wind maxima at the different stations suggests, however, that they tend to be lower at the mountain stations as compared with Boulder (i.e., Southern Hills).

More objective means of comparing wind speeds at the stations are average hourly wind speed cross sections along the line of stations. Since hourly data were available for three additional stations, it was possible to extend the line of 7 anemometer stations discussed above. Unfortunately, wind data for the D-1 station (a mountain station about 2 km east of the Divide, 2140 m above Boulder, and almost in line with the rest of the stations) were frequently missing, particularly wind direction, and the latter could not be interpolated from a nearby station (as was done in some cases for hourly data from the lower stations around Boulder where distances between stations are relatively short and the terrain more homogeneous). Wind data for the D-1 station were therefore not sufficiently available to be useful for this study. This left the Mines Peak station (at the Divide, about 30 km to the west of and 2200 m above Boulder) and Denver (about 35 km east of Boulder). Although both are about 25 km to the south of the main line of anemometer stations, they were used to extend the line to the east and west. Since the north-south extent of wind storms in the Boulder area seems to vary from storm to storm and especially since large variations in wind are possible in mountainous terrain, the results should be interpreted with these limitations in mind. Wind speeds for the Jones and 30th Street stations in Boulder, on the other hand, are not shown. Both stations are relatively close to the Southern Hills station which because of its open exposure is considered to be more representative of wind conditions at the foot of the mountains.

The cross sections shown in Figs. 9.23 to 9.27 represent distributions of average hourly wind speeds, west wind components and gusts for the time of wind storms in Boulder, the time of wind maxima at C-1 and also at Denver. They were computed by first determining the hour with the highest west wind component at the Southern Hills station during the 20 wind storms and three additional phases and similarly by determining the hour with the maximum westerly wind component at the C-1 and Denver stations. Hourly wind speeds as well as the west wind components and highest gusts for these hours were then averaged for all stations.

Regarding the time of the west wind component maximum, it was generally easier to pick this hour than might be expected from the 5-minute wind plots; furthermore, the hourly data always covered a longer period than the 5-minute data which was important for the determination of wind maxima at the mountain and Denver stations. As discussed above, wind speed at the mountain stations can sometimes be very low during times of wind storms in Boulder and since it was not considered to be very meaningful to include the maximum of such periods, an arbitrary threshold value of ≥ 8 m s^{-1} was adopted after examination of wind speeds at C-1 during the 20 wind storm cases. Besides the cases eliminated by this criterion, the group of 21 cases of west wind maxima at C-1 thus chosen was not equal in number to that for maxima at Southern Hills

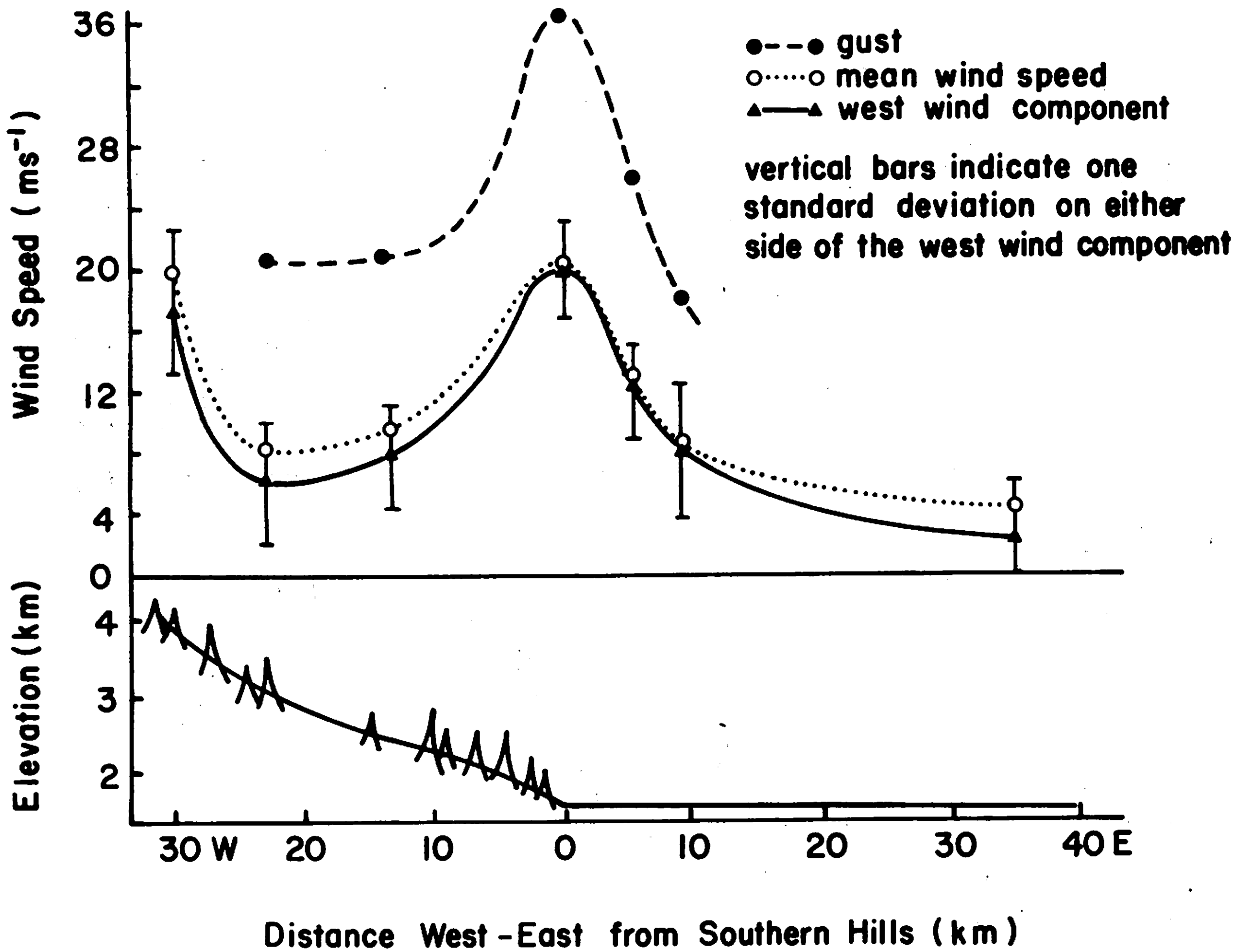


Fig. 9.23. Mean surface wind cross section for hour with highest west wind component at the Southern Hills station.

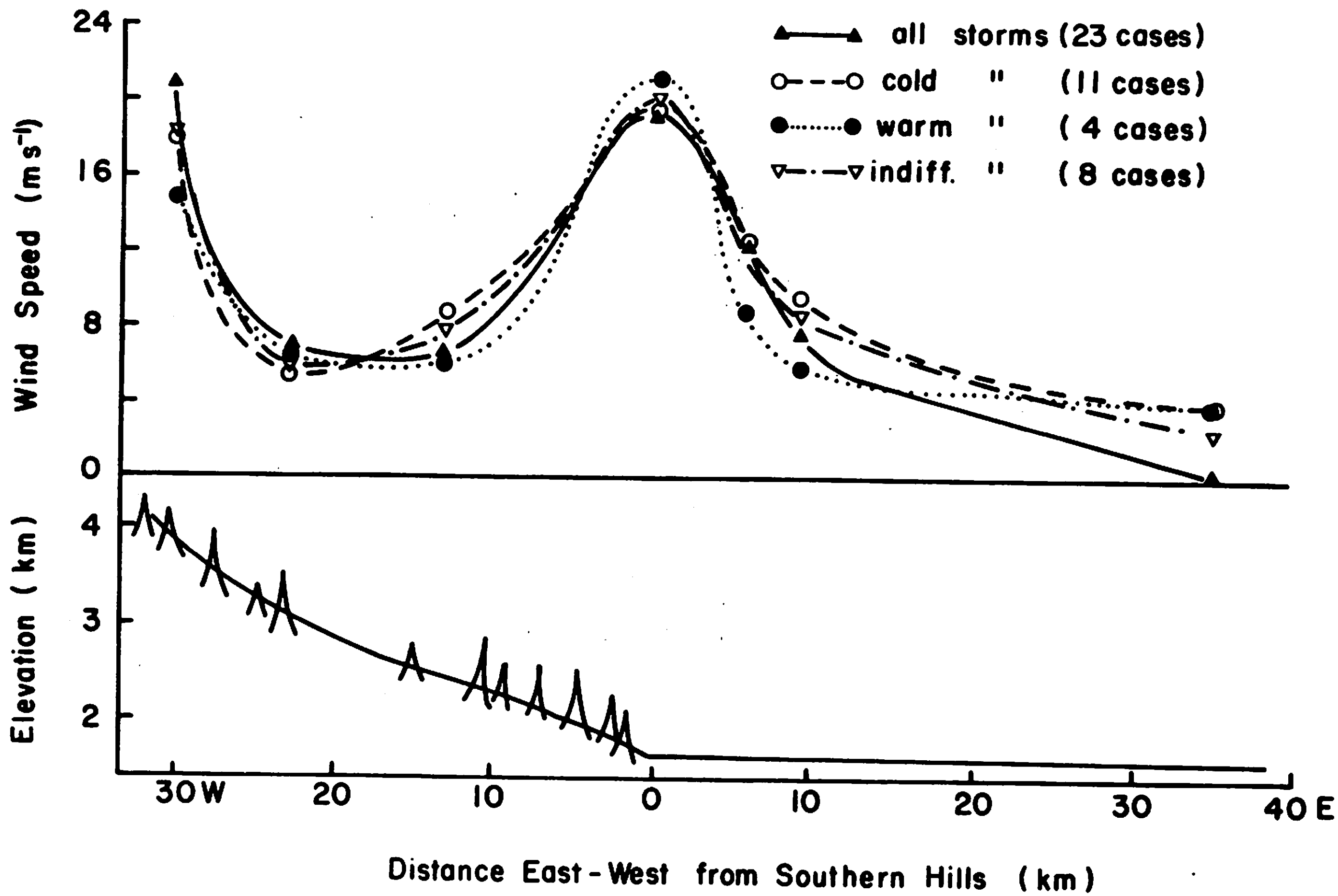


Fig. 9.24. Mean surface wind cross section for hour with highest west wind component at the Southern Hills station, classified by storm type.

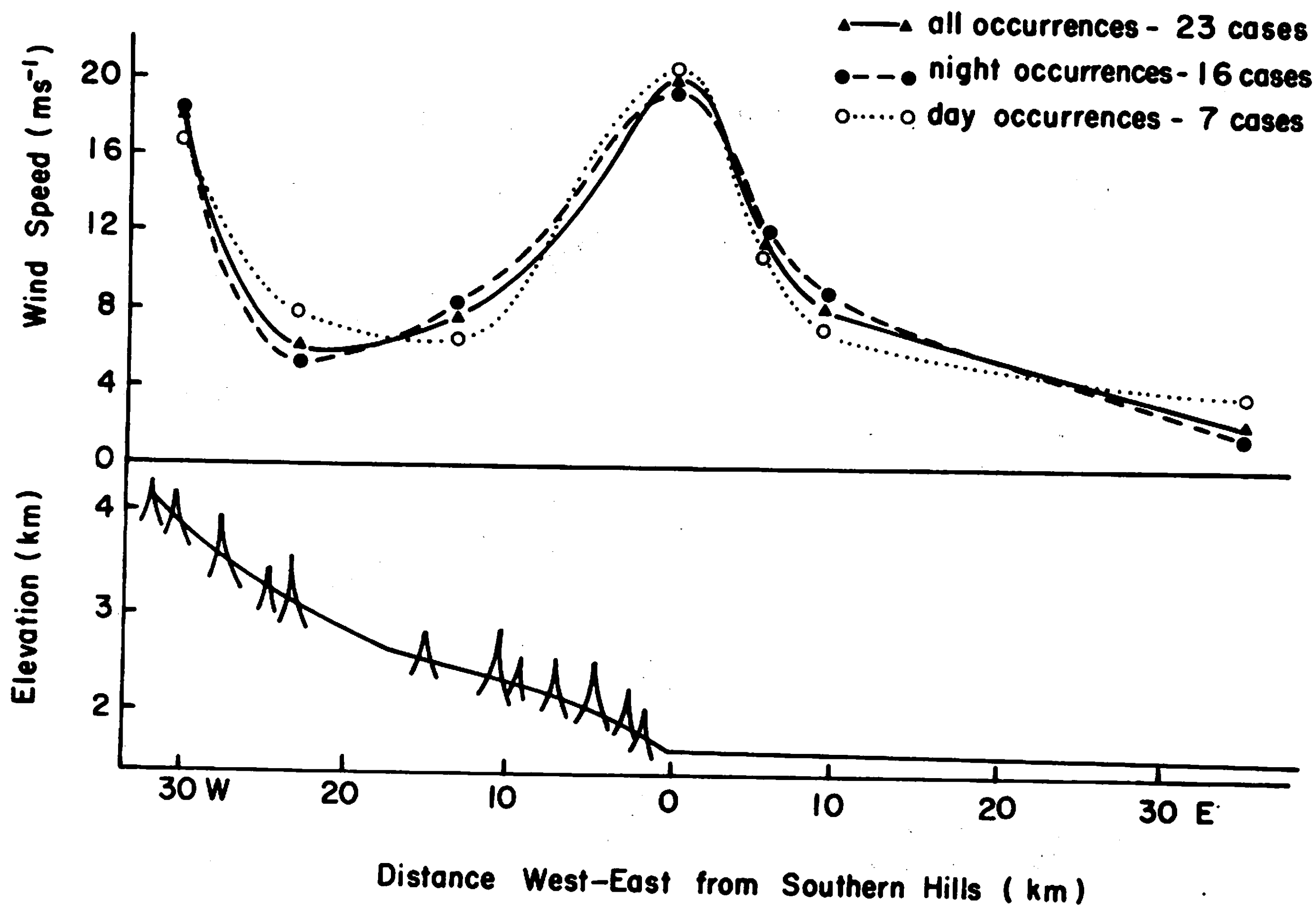


Fig. 9.25. Mean surface wind cross section for hour with highest west wind component at the Southern Hills station, classified by time of occurrence.

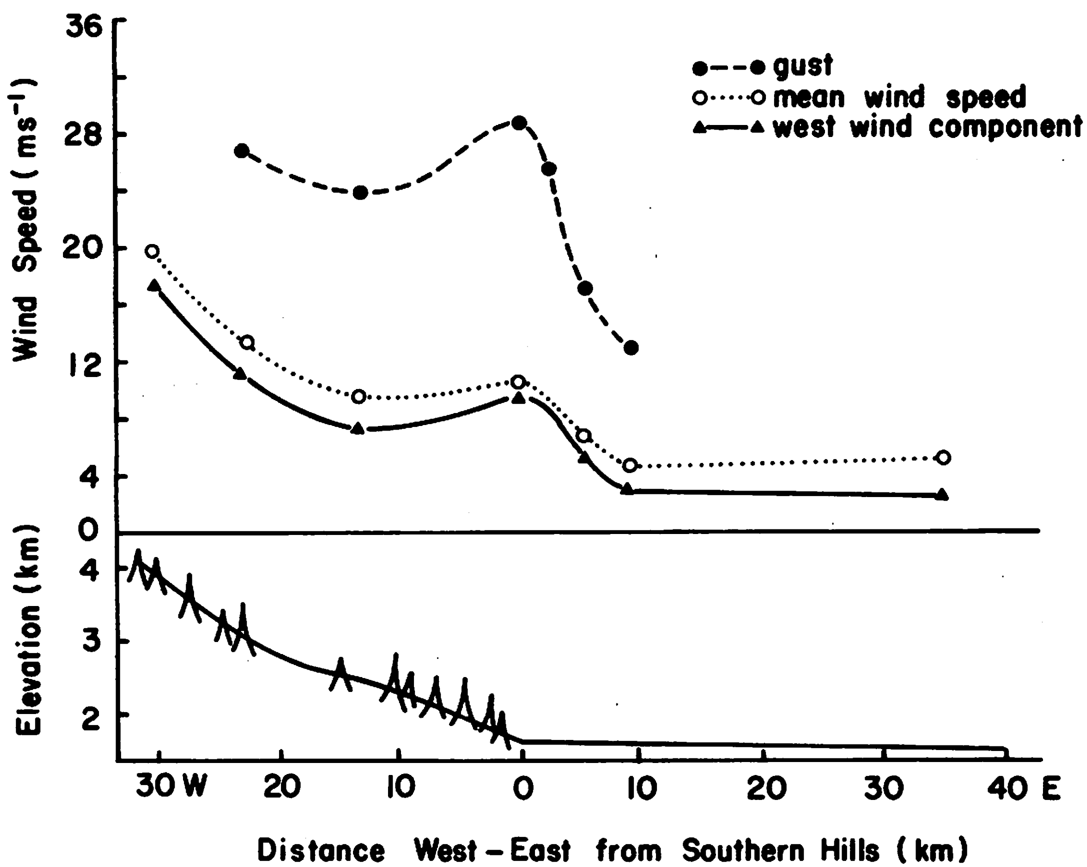


Fig. 9.26. Mean surface wind cross section for hour with highest west wind component at the C-1 station.

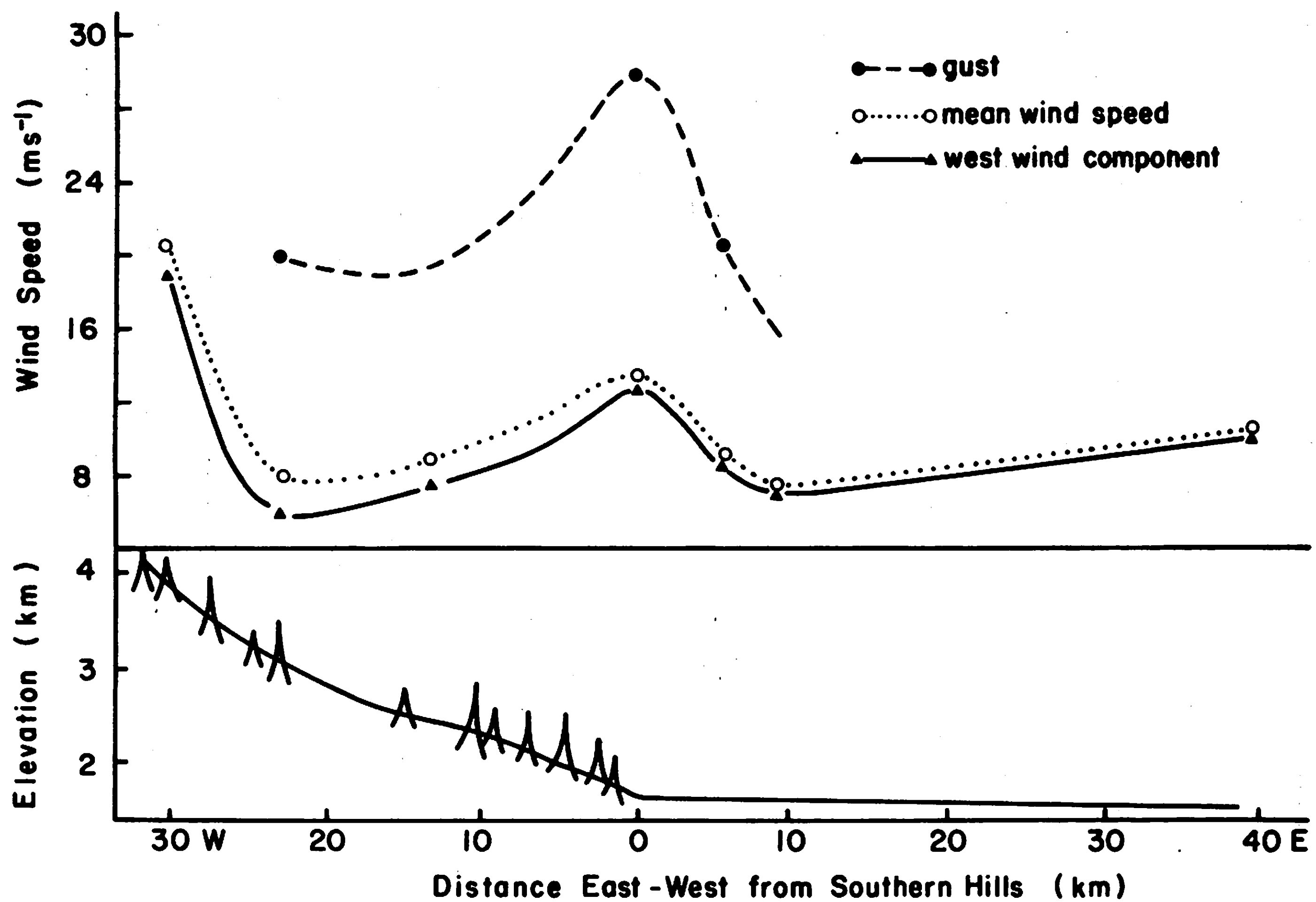


Fig. 9.27. Mean surface wind cross section for hour with highest west wind component at the Denver station.

because two phases of wind peaks at Boulder, separated by a pause, may be associated with only one maximum at C-1; and, vice versa, a storm maximum in Boulder may be preceded and/or followed by peaks at C-1, but these had to be separated by a distinct pause at C-1 in order to be included in the calculations as two maxima. For the Denver station the same criteria for defining maxima were used as for the C-1 station which reduced the number of cases for the Denver-based cross section to 19. In cases with more than one consecutive hour with the same maximum speed at C-1 or Denver, the hour closest to the time of the Boulder wind storm was used.

In considering the comparability of the wind data, anemometer exposure is probably the most important factor (see also Section 4). For all stations with heights greater than 3.4 m (the normal height for most of the instruments used in this study) the reduction necessary to normalize the data was estimated in Section 4. Effects on wind direction because of channeling or deflection is another factor to be considered since it would in turn affect the calculated west wind component. To estimate this effect, relative frequencies of wind direction for 5-minute mean speeds ≥ 15 m s^{-1} were computed for each station and are shown in Fig. 9.28. The cosine of the angle between 270° and the modal direction of high wind speeds (as given by the peak in the distribution for that station) was taken as the deflection factor, assuming that strong downslope winds are westerly. The tendency for the wind direction in the free air in the lowest layer to be about 10° north of west, discussed in Section 13, is of little significance here since it would change the estimated deflection factor only by 2 per cent.

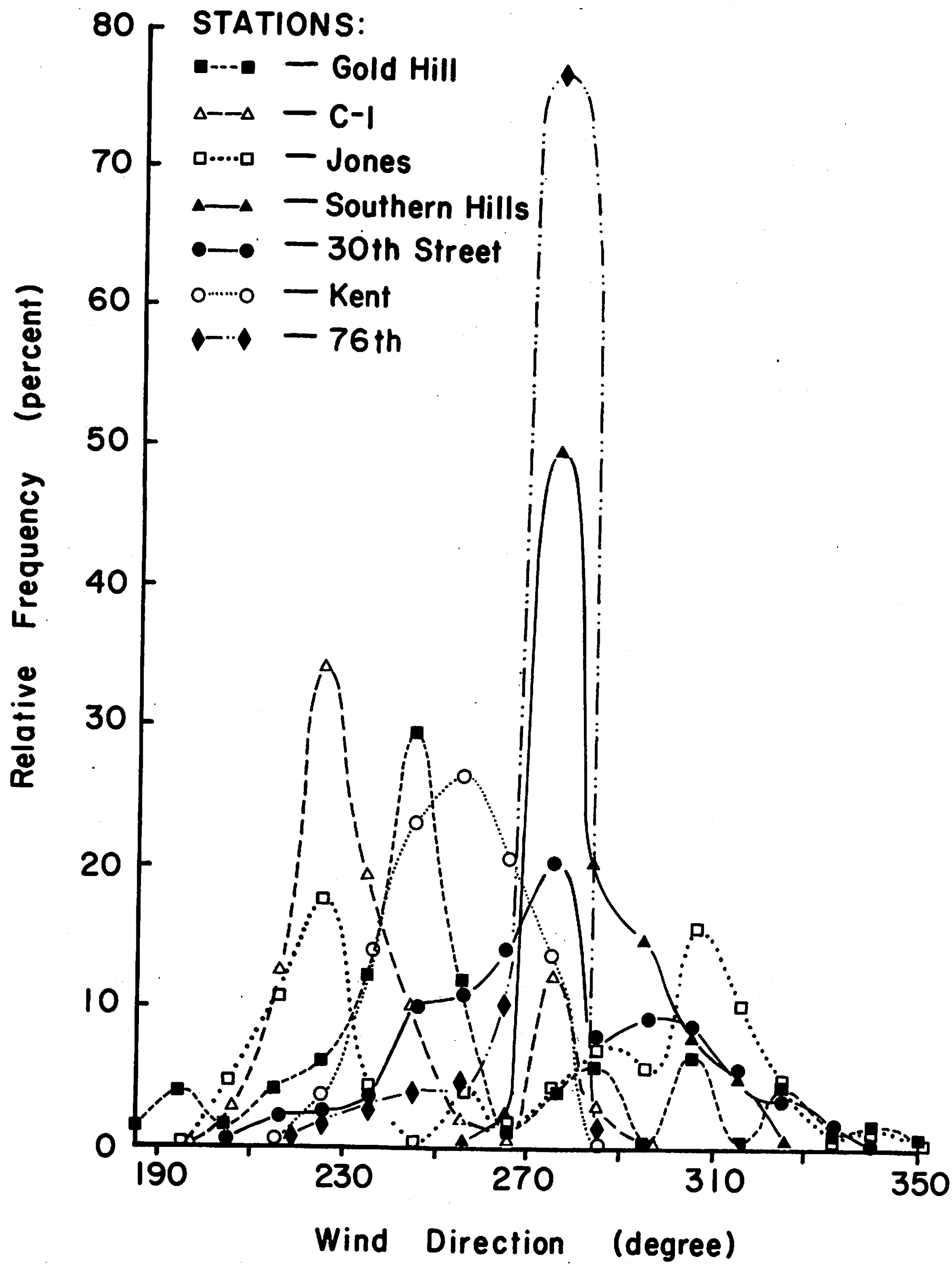


Fig. 9.28. Relative frequency distribution of wind direction for 5-minute mean wind speeds ≥ 15 m s^{-1} by 10 degree intervals.

The two factors, anemometer height and deflection, are compared in Table 9.1. The question mark for the Jones station refers to the interesting bimodal direction distribution; it points out the peculiar exposure of this station which will be discussed further in Section 10. For the Denver and Mines Peak stations open exposure was assumed. With the exception of the C-1 and Mines Peak stations the table shows that the effect of the two factors is small considering the wind speed range dealt with in this study. For Mines Peak both the wind speed and the west wind component should be reduced when comparing them with the rest of the stations. At C-1 strong winds obviously experience a significant deflection to the southwest considering the fact that wind direction in the free air over Denver during wind storms tends to be slightly north of west. Subtracting the effect due to anemometer height, the west wind component at this station may be too low by about 30 per cent.

The mean surface wind cross section for the time of the wind storm in Boulder is shown in Fig. 9.23. Its dominant feature is <u>the large peak at the foot of the mountains, in Boulder, and the rapid decrease in speed to the east</u>. Hourly west wind components of surface wind at the Southern Hills station during the height of the 20 wind storms and three additional phases averaged 18-20 m s^{-1} (considering the exposure factor). For the same hour the wind speed at the mountain stations was less than half this value, i.e., at Gold Hill the mean west wind component was 8-9 m s^{-1} and at C-1 only 6-8 m s^{-1}. To the east of Boulder there is a similar rapid decrease in speed: at the 76th Street station the

Table 9.1. Summary of effects of differences in anemometer heights and wind direction deflection

Station	Height factor	Deflection factor
Mines Peak	+ 17%	0%
C-1	+ 11%	- 41%
Gold Hill	0%	- 10%
Jones	0%	?
Southern Hills	+ 9%	0%
30th Street	+ 8%	0%
Kent	0%	- 3%
76th Street	0%	0%
Denver	+ 13%	0%

average hourly west wind component was 8 m s^{-1} and only 3 m s^{-1} at Denver. For comparison purposes, it might be of interest to note that the mean wind speed for January (the windiest month) was 6 m s^{-1} at C-1 for the years 1967 to 1970, 4 m s^{-1} at Gold Hill for the years 1967 to 1969 (Barry 1972) and 4 m s^{-1} at Denver for the years 1967 to 1970 (calculated from 'Local Climatological Data', U.S. Department of Commerce). This means that during wind storms in Boulder speeds at C-1 and Denver were slightly higher than normal but practically double at the Gold Hill station. The increase at the latter is therefore probably significant; still, it is only half the speed reached at the foot of the mountains. The fact that not only at the time of the wind maximum in Boulder but throughout the major storm period the wind at the slope stations does not tend to reach the same intensity, is also demonstrated in the diurnal frequency distributions presented in Section 7. These show that the number of high gusts and 5-minute mean speeds is much less at C-1 compared with Southern Hills (the anemometer height at the two stations is about the same and wind direction [and thus possible deflection] was not considered in the computations of the frequency distributions).

Because of the diversity in wind speed patterns, the standard deviations of the means for the stations are relatively large. However, the differences in the means for Southern Hills and Gold Hill on one side of the wind peak and for Southern Hills and 76th Street on the other side are both significant. The wind maximum at Southern Hills (i.e., at the foot of the mountains) can therefore be considered to be real. The rapid increase in wind speed between Gold Hill and Southern Hills and the sharp decrease in speed between Southern Hills and 76th Street are in agreement with the local pressure minimum at Boulder which is associated with the 'lee air flow disturbance' during storms (as discussed in Section 12).

After grouping the cases into cold, warm and indifferent wind storms (as defined in Section 8), as well as into daytime and nighttime occurrences (Figs. 9.24 and 9.25), t-tests of the differences in the means between each of these groups and the overall mean at each station, as well as between the means for the different groups, showed no significant differences. Out of 81 tests -- 9 possible combinations for 9 stations -- only one difference was found to be significant. Furthermore, since only two F-tests showed significant differences in variance, it may be assumed that the groups have come from the same population.

In Fig. 9.26 (representing the average cross section of westerly surface wind for the hour with a wind maximum at the C-1 station, exceeding 8 m s^{-1} by definition) the peak at Boulder has almost disappeared while the average speed at C-1 has increased to about 11-13 m s^{-1}. This is significantly higher than the mean speed of 6-8 m s^{-1} for the time of wind storms in Boulder. It shows that when a wind speed maximum at the C-1 station occurs, it does not tend to coincide with the wind storm peak in Boulder which supports the hypothesis of the 'moving wind maximum'. But the average speed of 11-13 m s^{-1} at C-1 is still significantly less than the average of 18-20 m s^{-1} at Southern Hills at the height of the wind storms in Boulder.

Since the difference in time between the occurrence of wind maxima at C-1 and Southern Hills varied between about zero and 10 hours, the average wind speed cross section shown in Fig. 9.26 includes hours during which the wind speed was also at a maximum at Southern Hills. This probably accounts for the small hump at the Boulder station and the significant decrease to the east. Again F- and t-tests showed little significant difference in the means between the cold, warm and indifferent cases, and between daytime and nighttime occurrences.

The average wind speed cross section for the time of the maximum at Denver, exceeding 8 m s^{-1} by definition, is presented in Fig. 9.27. The mean west wind component at that station, about

10 $m\ s^{-1}$, is significantly higher than 3 $m\ s^{-1}$ reached at the time of the wind storm in Boulder. As in the case of the mountain station maxima, this indicates that when a wind maximum at Denver occurs, it does not tend to coincide with wind storms in Boulder. Another similarity is the still significantly lower speed at Denver compared with the mean of 18-20 $m\ s^{-1}$ at Southern Hills during wind storms in Boulder. It is known, however, that severe and damaging winds do occasionally occur in Denver (as well as on the slopes); but, aside from such less frequent occurrences (particularly at Denver), the criteria used for the selection of the time periods for this study were based on wind speeds in Boulder and therefore probably do not include some occurrences of high winds at the slope stations (or Denver) if they were not associated with wind storms in Boulder.

The difference in time between the occurrence of wind maxima in Denver and Boulder again varied between zero and 10 hours which explains the small but still significant wind speed peak at Southern Hills. Cold, warm and indifferent cases as well as daytime and nighttime occurrences, were also found not to be significantly different.

Summarizing the results, although there is much complexity in wind speed patterns for the storms in general as well as at the individual stations, there are some outstanding common features:

1. the wind storms are extremely localized phenomena. Wind speeds increase by about 50 per cent over a distance of 10 km between the mountain stations and Boulder, highest speeds are reached at the foot of the mountains in Boulder,

and wind speeds decrease again rapidly by 50 per cent over a distance of about 10 km to the east into the plains;

2. wind speed maxima in the mountains to the west and on the plains to the east are lower than those at the foot of the mountains. They also tend not to occur simultaneously as a result of the non-stationary nature of the phenomenon. Consequently, Boulder wind storms were found not to be associated with similarly strong winds on the slopes as would be expected from the connotation of the word 'downslope storm';

3. wind storms with different temperature characteristics (i.e., warm, cold, indifferent), as well as daytime and nighttime occurrences, were found not to differ in terms of surface wind cross sections.

10. Turbulence Characteristics

In addition to the hurricane-force speeds, extreme gustiness is another outstanding characteristic of Boulder's wind storms. Anemometer traces of the three most severe storms over the last 6 years have been presented by Kuettner and Lilly (1968), Julian and Julian (1969) and Lilly and Zipser (1972). Particularly the traces shown in the first two references (recorded on the roof of the NCAR building, about 23 m above ground) demonstrate well the very turbulent nature of the wind, with speeds increasing from less than 5 m s^{-1} to 45 m s^{-1} in a matter of seconds. Figs. 10.1 and 10.2 are anemometer traces from 4 different stations for two of the 20 wind storm cases; the 4 traces are arranged in the same

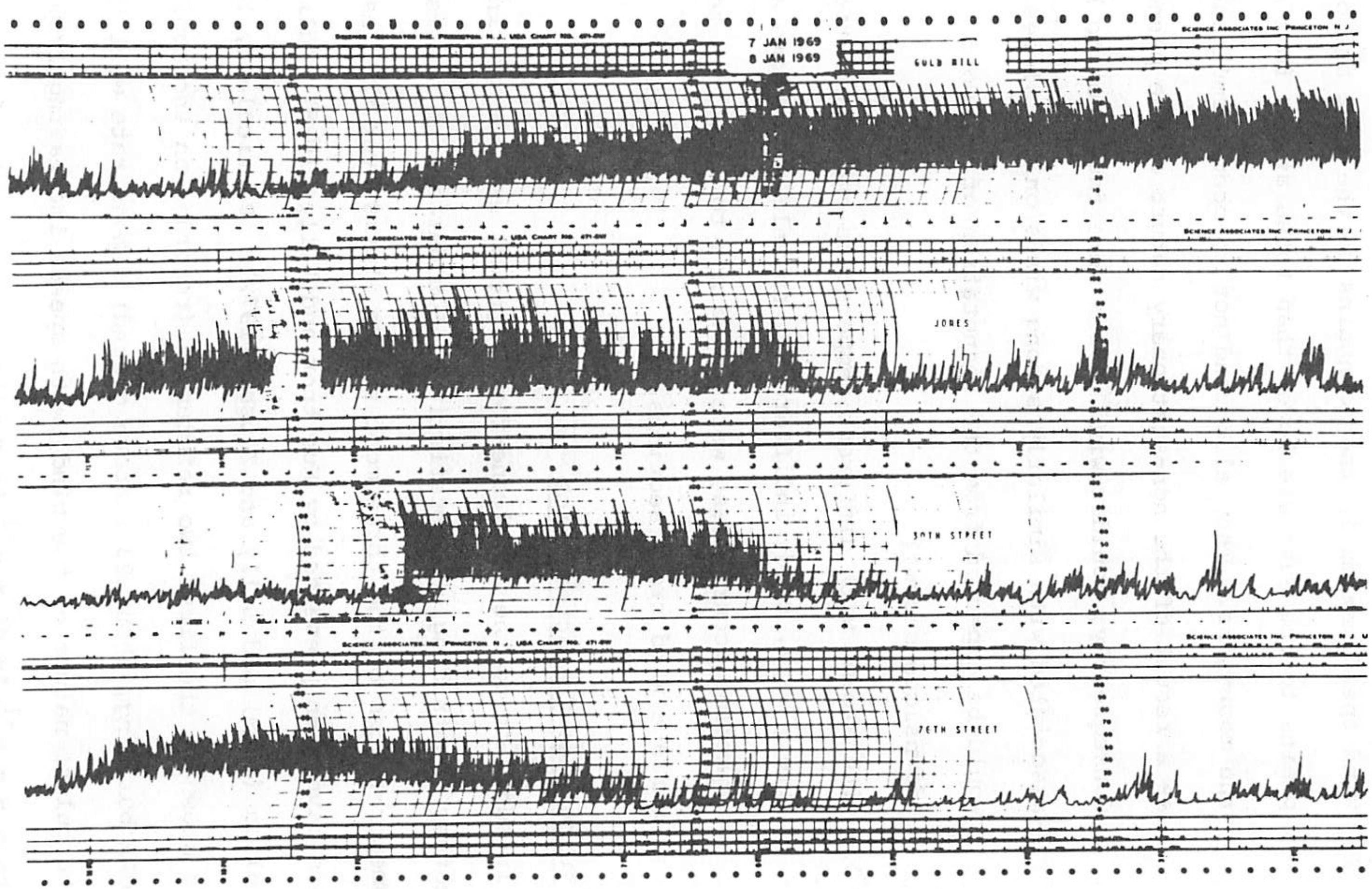


Fig. 10.1. Anemometer traces from the east slopes of the Front Range on 7 January 1969 (vertical scale is in miles per hour, time is read from right to left; a power failure stopped the chart drive at the Jones and 30th Street stations and the latter was not restarted until after the storm).

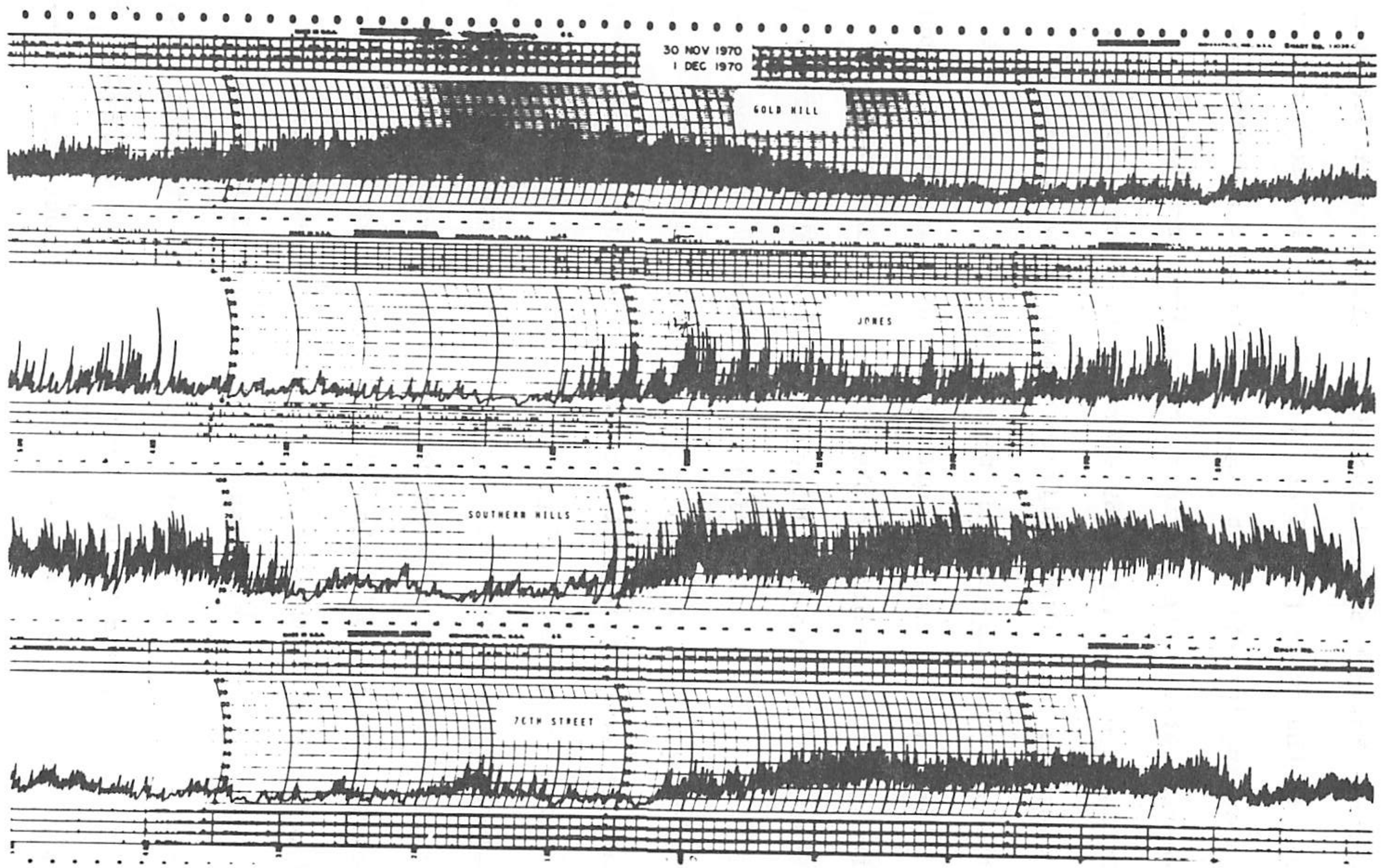


Fig. 10.2. Anemometer traces from the east slopes of the Front Range on 30 November-1 December 1970 (vertical scale is in miles per hour, time is read from right to left).

way as the computer plots shown in Section 9, with the top trace representing the highest mountain station in the diagram. The two figures show not only the striking difference in gust structure between (1) the storm period and (2) the time before and after it as well as during 'pauses', but also illustrates some interesting differences between the stations, to be discussed below.

During the 7 January 1969 storm approximately 50 per cent of the residences in Boulder suffered damage (City of Boulder 1970). Both the high mean speeds and, perhaps more so, the extreme gusts of Boulder's winds, together with the associated flying debris, are directly responsible for the severity of this damage. Illustrations presented by Julian and Julian (1969), Lovill (1969) and Lilly and Zipser (1972) indicate the enormous forces involved. Lilly and Zipser (1972) point out, however, that Boulder's altitude reduces the damage potential from static loading slightly since the air density is about 15 per cent less than at sea level. This corresponds to an 8 per cent decrease in wind speed.

While the action of wind has been of concern to the structural engineer, there is no simple rule for indicating the severity. For a general guide the gust factor has been used. This is defined as the ratio of the peak gust to the mean speed for a given averaging period which is usually 5 minutes (Lettau and Haugen 1960). As summarized by Davis and Newstein (1968), results from a number of investigations have shown that the gust factor decreases with increasing wind speed, because of a slower increase in gust as compared to the mean speed. However, there may not be any real variation of the gust factor with mean speeds exceeding about

15 m s^{-1} (Lettau and Haugen 1960). Also, the factor will increase with increasing averaging time, because of increased probability of a high gust; and it also tends to increase with decreasing anemometer height, because peak gusts decrease downward less rapidly than mean speeds (Deacon 1955). The sensitivity of the anemometer is another consideration since a sluggish instrument will not record the highest gust and at the same time will overestimate the mean speed. Considering the large distance constant of the instruments used in the present study (7-second gusts are resolved with an accuracy of 90 per cent), it may be assumed that the instantaneous gusts and thus the gust factors were higher than those recorded.

To evaluate the gustiness of Boulder's wind storms, average and maximum gust factors for all 20 wind storm cases were calculated for the 7 anemometer stations in and around Boulder for which 5-minute data was abstracted. These stations all have the same type of instrument and are thus comparable, except for anemometer height. Figs. 10.3 and 10.4 show the results as a function of the 5-minute mean speed. Mean gust factors calculated from less than 5 occurrences (i.e., for very high mean speeds) were discarded because of increasing variability. Gustiness components along the x-, y- and z-axis could not be calculated because of the discontinuous (once a minute) wind direction records and the limitation of the direction resolution to 8 compass points.

In general, the curves in Figs. 10.3 and 10.4 show the expected decrease in the mean and maximum gust factors with increasing mean speed. Apart from some local variations (including differences in

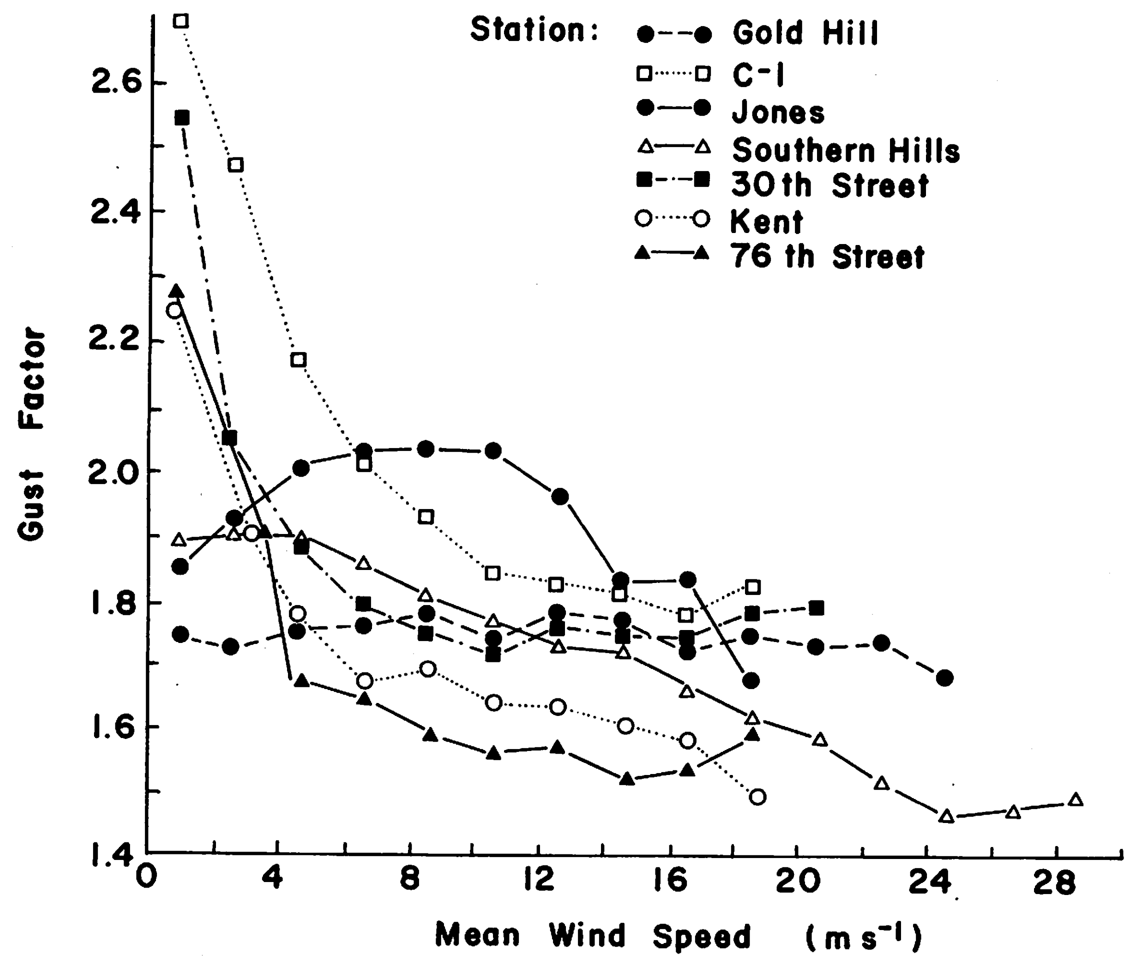


Fig. 10.3. Mean gust factor for 5-minute mean speeds.

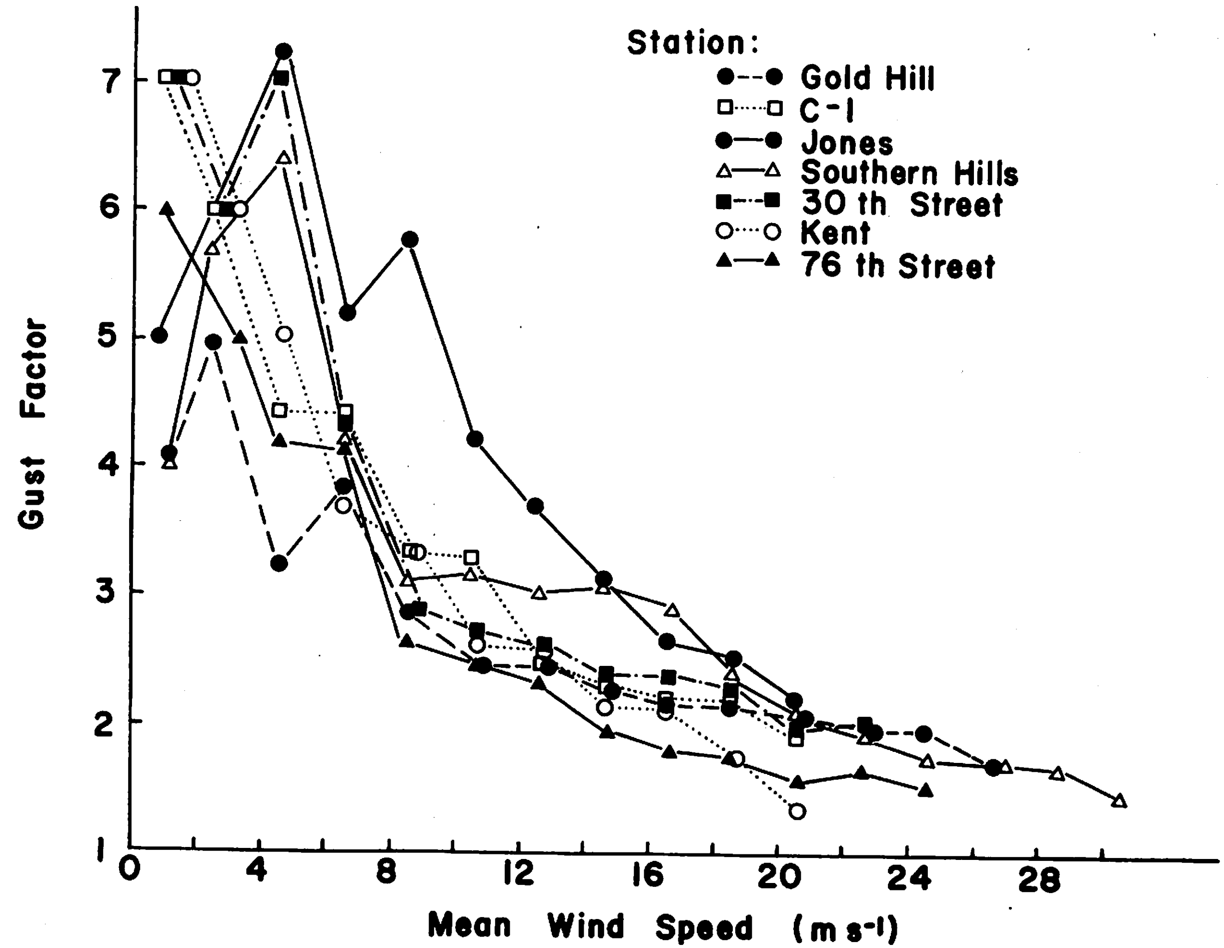


Fig. 10.4. Maximum gust factors for 5-minute mean speeds.

anemometer heights) a general trend of decreasing gustiness with increasing distance from the foot of the mountains is indicated. It was shown in Section 9 that mean speeds decrease with distance from the mountains; the decrease in the gust factor with increasing distance must therefore be the result of a decrease in gusts at a faster rate than the decrease in the mean speed.

At the Jones location the gust structure is rather unusual; the gust factor actually increases with increasing mean speed up to about 12 m s^{-1} and the maximum gust factor is generally higher than at any of the other stations. Comparison of the anemometer traces for Jones and the Southern Hills station to the east of it, or even the 30th Street station to the northeast, (Figs. 10.1 and 10.2) shows that the difference is due to similar gusts but lower mean speeds at the Jones station. Only 8 to 9 per cent of the higher mean speeds at 30th Street and Southern Hills can be explained if an adjustment is made for difference in anemometer levels. The gust structure of the wind at Jones is probably the result of its location on sloping ground below the almost vertical, sharp-edged Flatirons causing separation and lee eddies.

At Gold Hill the relatively constant mean and low maximum gust factors may be the result of the surrounding aspen trees. But since these are leafless during the winter, a more important factor may be its very open exposure to the west (situated on a ridge), making it more representative of conditions in the free atmosphere in which turbulence is usually less.

Lettau and Haugen (1960) have calculated mean and maximum gust factors, using data from a number of sources and assuming that differences of instrumentation and exposure heights do not affect the representativeness in terms of average gustiness conditions. Their results are shown in Fig. 10.5 together with mean and maximum gust factors for all 7 Boulder area anemometers combined. The differences clearly demonstrate the much higher than average gustiness of the Boulder wind storms.

Many of the downslope winds occurring in different parts of the world are described as extremely gusty, and anemometer traces for downslope winds in the lee of the Pennines, Great Britain, and on Hokkaido, Japan, as well as for a Foehn at Altdorf, Switzerland, and a Bora at Dubrovnik, Yugoslavia, are shown in Section 2 (Figs. 2.1 to 2.3 and 2.5). These traces only permitted the abstraction of gust and mean speeds for one-hour averaging periods. The gust factors for these 4 cases are shown in Fig. 10.6 together with mean gust factors for the Boulder area (recalculated for the same averaging period, as shown in Fig. 10.7). Considering the differences in instrumentation, anemometer height and the small sample size from the 4 areas, the gustiness characteristics of Boulder's storms are very similar to those of other bora-type and foehn-type winds around the world.

Another approach to the study of the turbulence of Boulder's winds is based on Reynolds' classic theory which regards turbulent flow as comprising of a steady, or slowly varying, mean flow on which eddies are superimposed (i.e., $u = \bar{u} + u'$). The energy distribution among eddies of different scales can be studied by means of spectrum analysis techniques which resolve a time

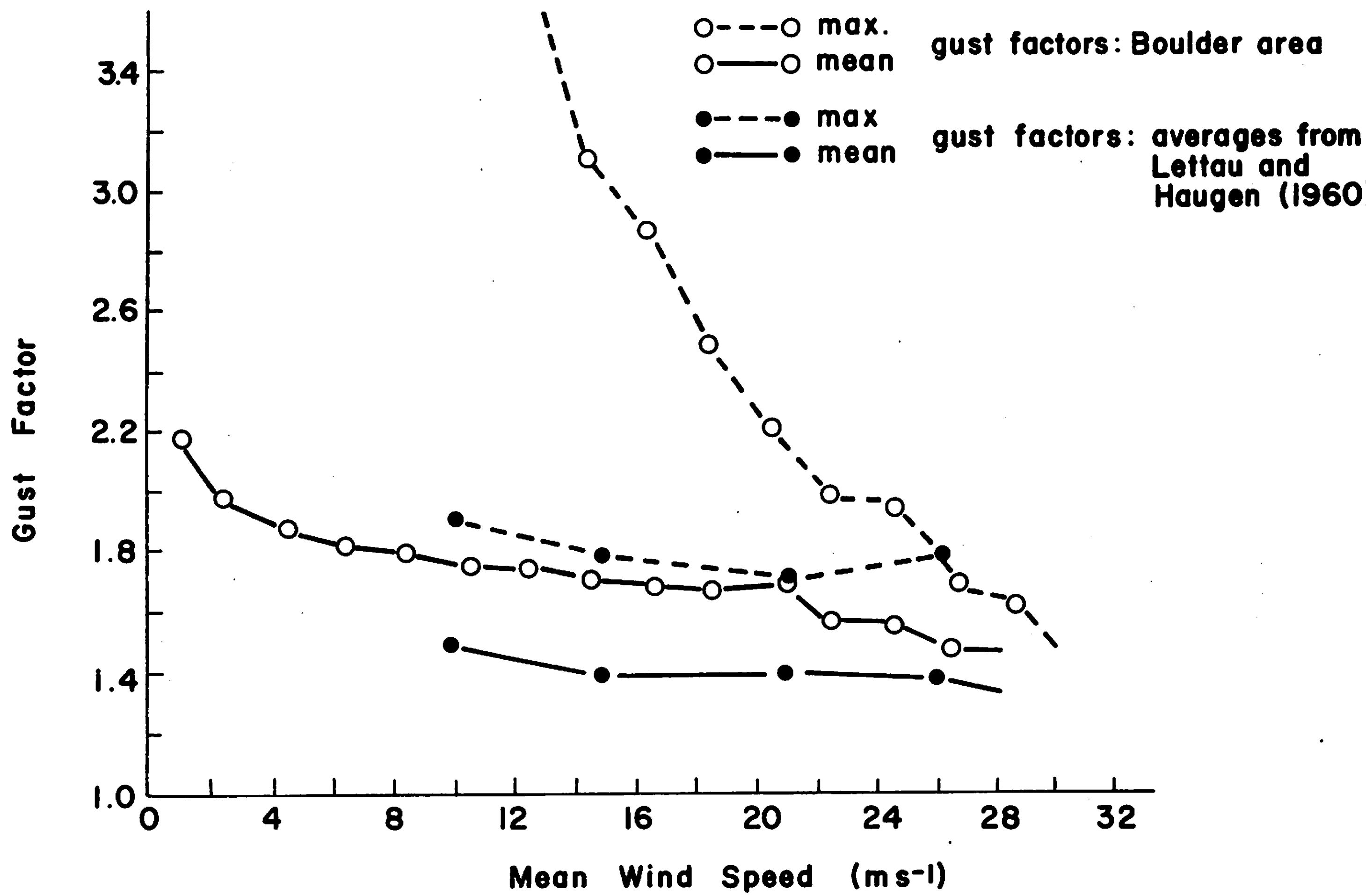


Fig. 10.5. Mean and maximum gust factors for 5-minute mean speeds.

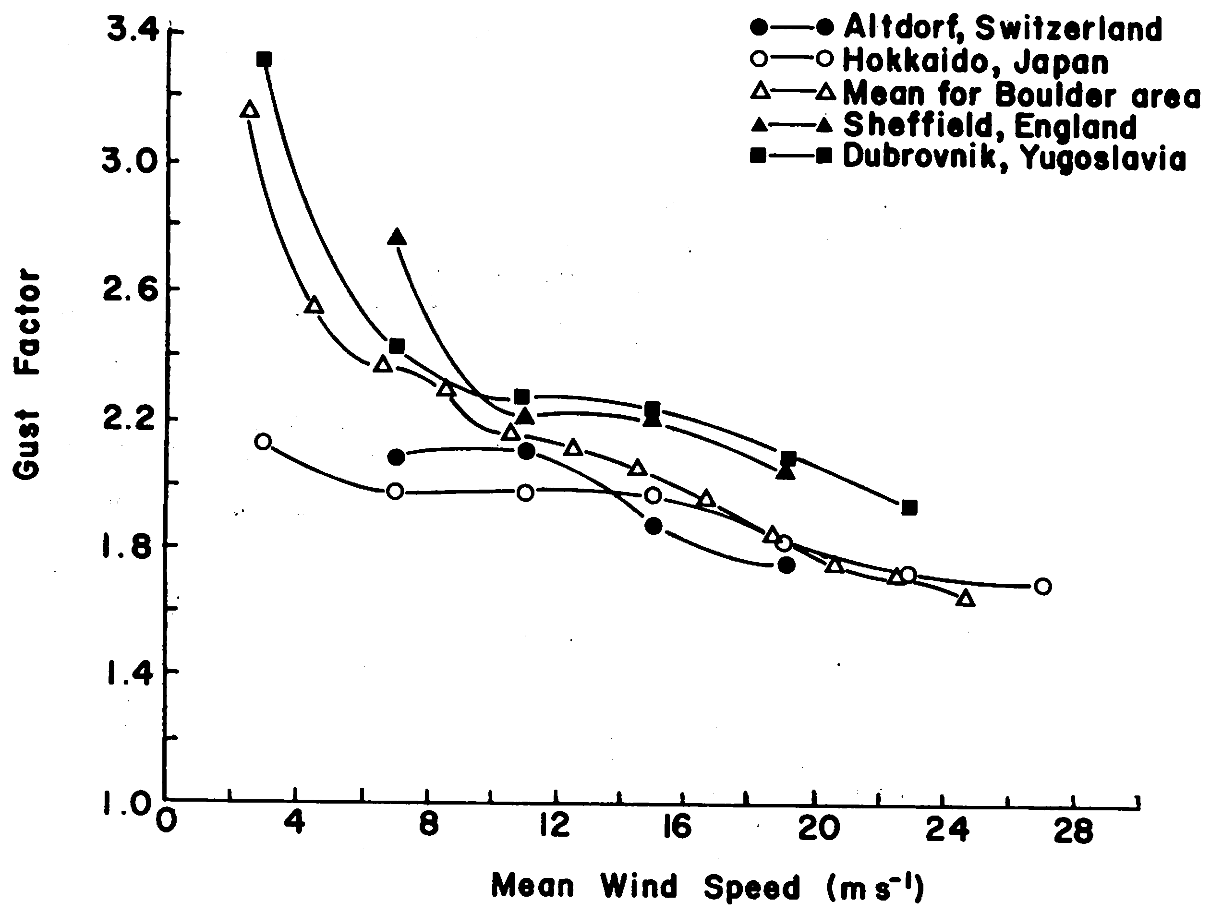


Fig. 10.6. Mean gust factors for one-hour mean speeds for 5 downslope wind storm locations around the world.

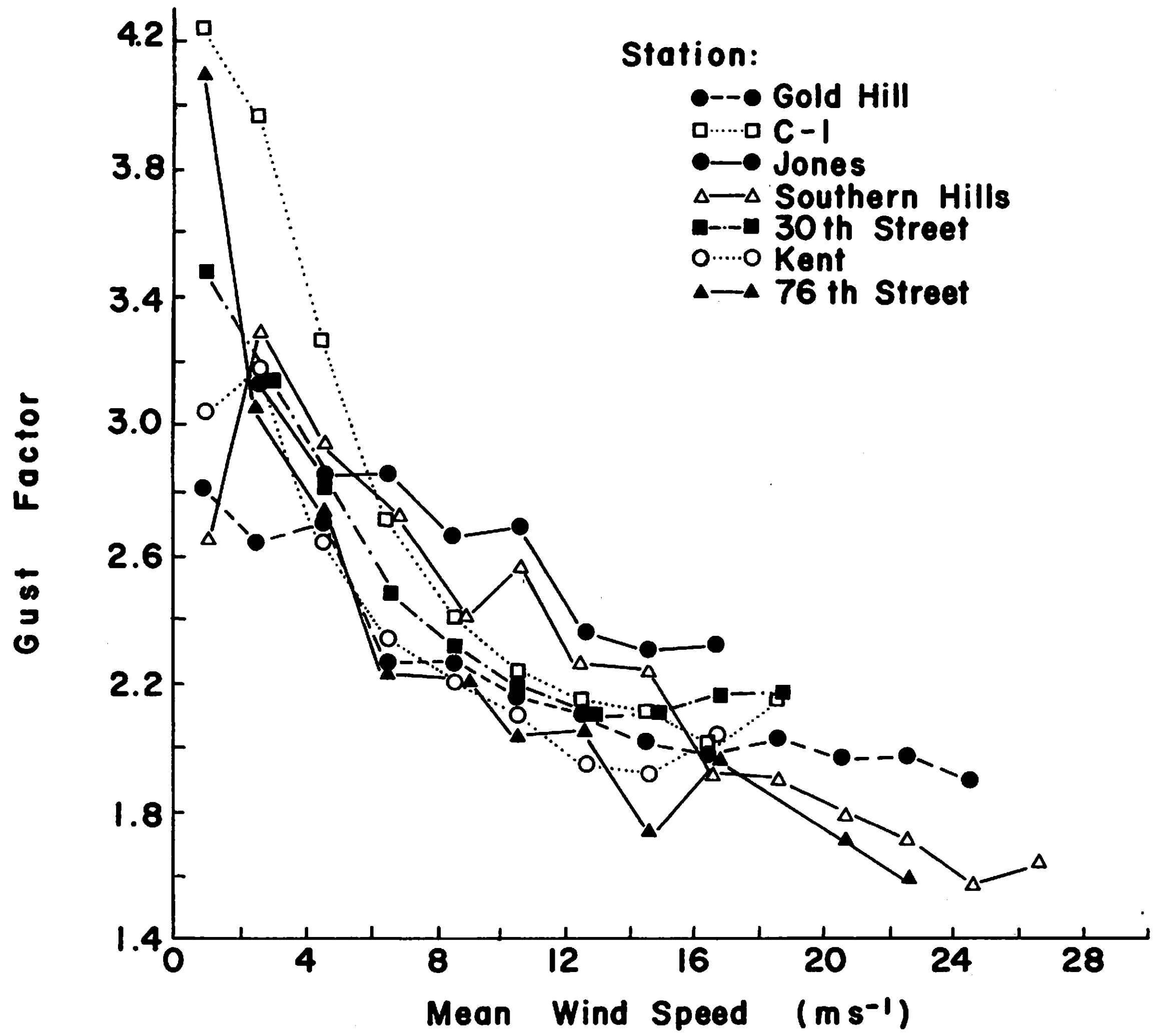


Fig. 10.7. Mean gust factors for one-hour mean speeds for the Boulder stations.

series into a series of harmonic components of different wave lengths.

Spectral analyses of horizontal wind speed have shown the existence of two major eddy-energy peaks in the spectrum, one at the synoptic scale, at a period of about 4 days (caused by migratory pressure systems), and the other at the microscale, at a period of about one minute (caused by mechanical and convective type of turbulence) (Fig. 10.8). The spectral gap in the range between periods of about 10 minutes to two hours has been found under varying terrain and synoptic conditions, as well as in conditions of strong winds and is thought to be due to the absence of physical processes supporting wind speed fluctuations in this range (Panofsky and Van der Hoven 1955; Van der Hoven 1957; Davenport 1961; Fiedler and Panofsky 1970). The existence of this spectral gap, however, should still be regarded as uncertain. Tyson (1968), for instance, found that the nocturnal mountain wind contributes its maximum energy at a period of about one hour.

Five-minute gusts and mean wind speeds from the 7 anemometer stations in and around Boulder for a number of wind storms were subjected to spectral and cross-spectral analyses. The appearance of a 'significant' peak in the frequency range covered by the available data (i.e., between 0.5 and 0.1 cycles per 5 minutes, corresponding to periods of 10 minutes to about one hour) is, however, unlikely, since it is within the range in which the spectral gap is located. Moreover, surges of the type found by Tyson (1968), due to pressure gradients created by the greater nocturnal cooling along the slope than in the free air at the

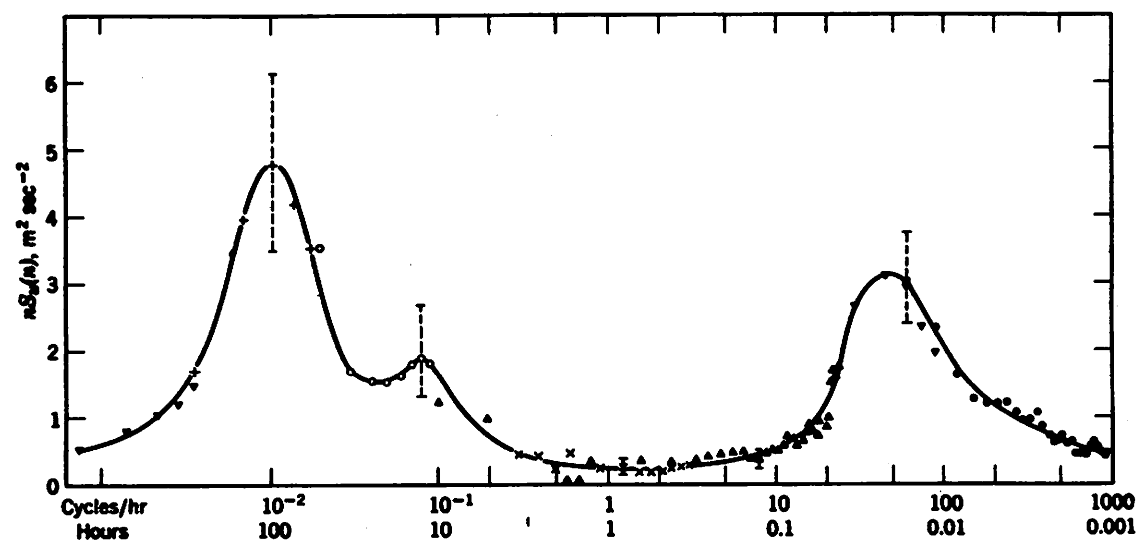


Fig. 10.8. Schematic spectrum of wind speed near the ground estimated from a study of Van der Hoven (1957).

same height (Fleagle 1950), would not be supported by the strong winds during Boulder's storms. Because of the (not unexpected) negative results, only a few of the spectral diagrams are discussed. The method of computation was that described by Madden and Julian (1971).

Figs. 10.9, 10.10 and 10.11 show the gust spectrum for the 30 November-1 December 1970 storm at the three stations Jones, Southern Hills and Kent (since they are located relatively close together and along an almost east-west line). To test the significance of the spectral peaks a 'null' continuum has to be fitted to the spectrum. Following Mitchell's (1966) outline for constructing such a continuum it was found that the spectra consisted of white noise (indicating an uncorrelated noise) and red noise (indicating persistence). This is not unusual and a background continuum, an average curve through the spectrum, is frequently assumed (e.g., Madden and Julian 1971). The selection of the confidence limits was based on _a posteriori_ significance theory since there is no _a priori_ hypothesis to explain periodicities in the frequency range covered by the available data. As discussed in detail by Mitchell (1966), a more stringent criterion is required to judge the statistical significance than if the spectral estimates in question correspond to an _a priori_ model. One method would be to raise the confidence level. From tabulations published by Mitchell, the 0.1 per cent confidence limits for 13 degrees of freedom (a function of the sample size and the bandwidth used in the spectrum analysis) are 2.69 and 0.38 times the local background continuum. These limits together with the

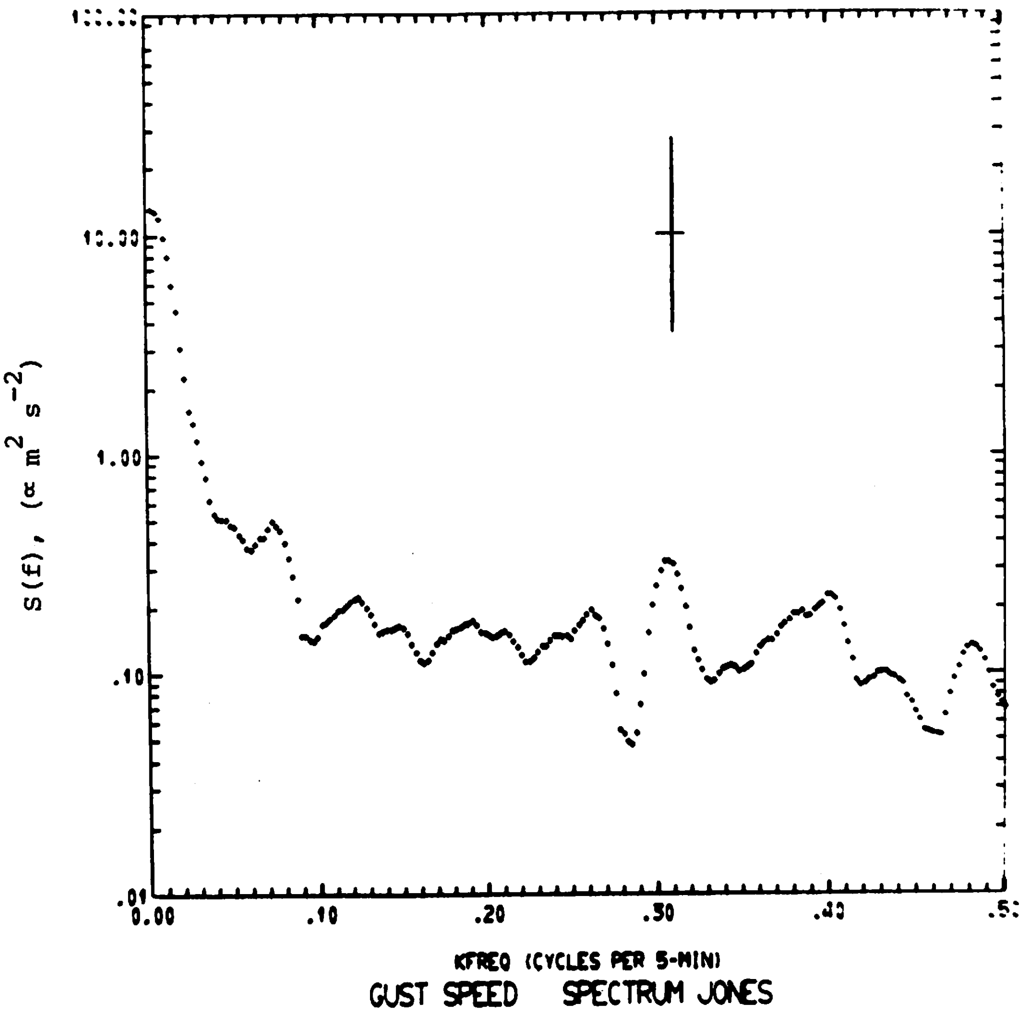


Fig. 10.9. Gust spectrum at Jones station for the 30 November-1 December 1970 storm. (The bandwidth used in the spectrum analysis and the 0.1 per cent confidence limits are shown by the cross. The logarithmic ordinate permits the cross to be moved anywhere in the diagram to determine the significance of individual peaks.)

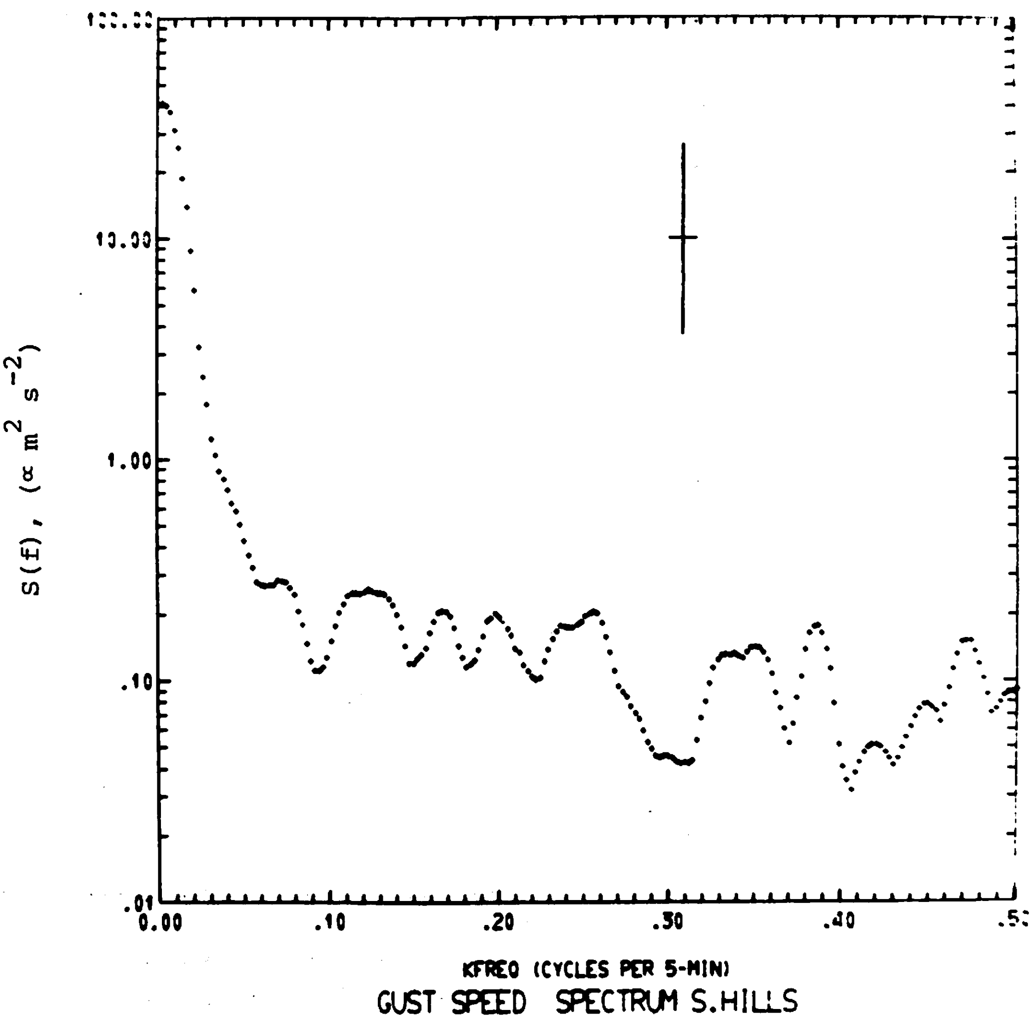


Fig. 10.10. Gust spectrum at Southern Hills for the 30 November-1 December 1970 storm. (The bandwidth used in the spectrum analysis and the 0.1 per cent confidence limits are shown by the cross. The logarithmic ordinate permits the cross to be moved anywhere in the diagram to determine the significance of individual peaks.)

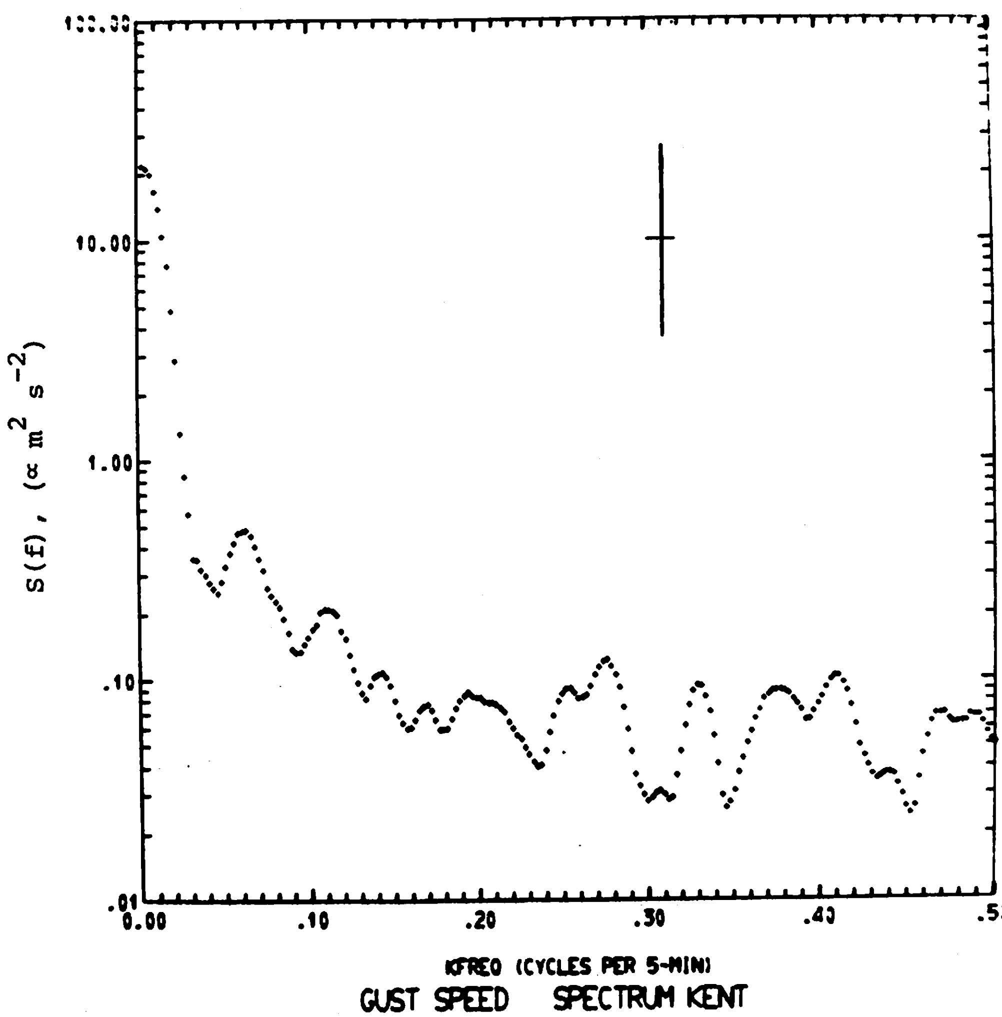


Fig. 10.11. Gust spectrum at Kent station for the 30 November-1 December 1970 storm. (The bandwidth used in the spectrum analysis and the 0.1 per cent confidence limits are shown by the cross. The logarithmic ordinate permits the cross to be moved anywhere in the diagram to determine the significance of individual peaks.)

bandwidth are shown by the cross in the spectral diagrams. The use of a logarithmic ordinate permits a constant scaling so that the cross may be moved anywhere in the diagram to determine the significance of individual peaks. Clearly, none of the peaks in Figs. 10.9, 10.10 and 10.11 are significant except for the low-frequency portion of the spectra.

Figs. 10.12 and 10.13 show the 'coherence-squared' statistic between the stations Jones/Southern Hills and Southern Hills/Kent (this statistic is analogous to the square of the correlation coefficient between frequencies, discussed in detail by Jenkins and Watts (1968)). For 13 degrees of freedom a 'coherence-squared' of 0.56 is significant at the 0.1 per cent level. Again, only the low-frequency portion of the spectra show significant coherence.

Thus, for this and all other storms analysed, most of the kinetic energy at the stations and coherence between stations was always found in the low frequency range (periods of more than one hour) because the storms themselves are essentially large non-periodic surges lasting about 6 to 12 hours. At the higher frequencies (periods of between 10 minutes and one hour) the spectrum was found to be extremely flat for a storm such as the 28-29 November 1970 (Fig. 9.11) during which the size of the gusts varied little. During the 30 November-1 December 1970 storm the gusts varied considerably (Fig. 9.12). However, the examples of the spectra for this storm (Figs. 10.9, 10.10 and 10.11) show that the peaks are still non-significant. The energy in the non-resolvable high frequencies (periods of less than 10 minutes) are added to and cannot be distinguished from the resolved range

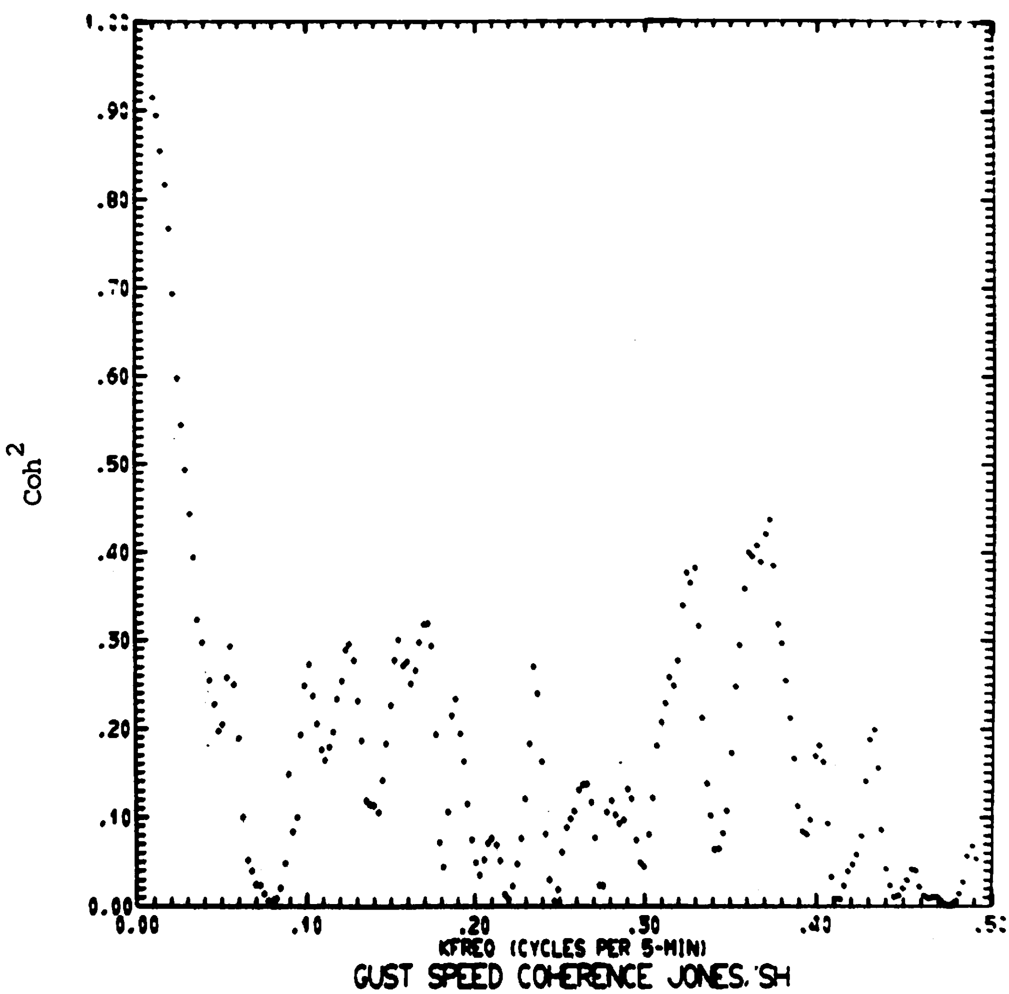


Fig. 10.12. Coherence statistic between gusts at Jones and Southern Hills for the 30 November-1 December 1970 storm.

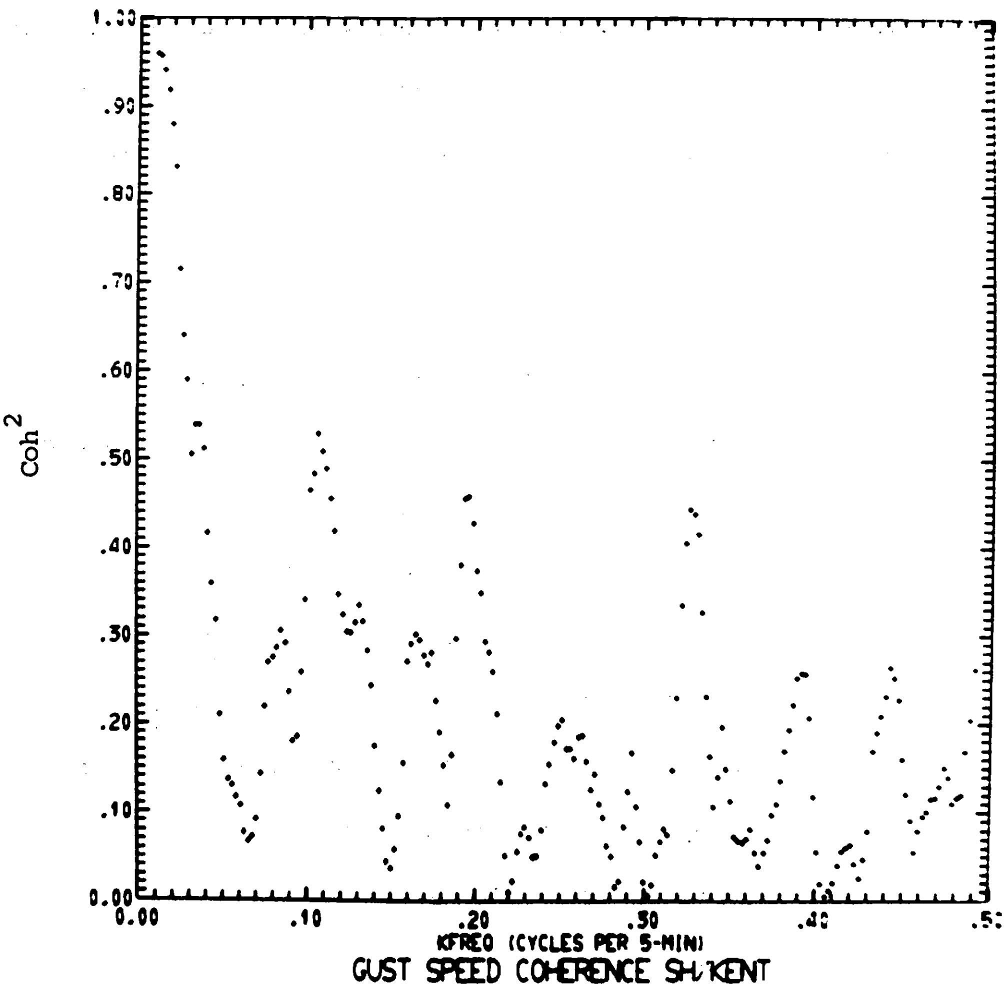


Fig. 10.13. Coherence statistic between gusts at Southern Hills and Kent for the 30 November-1 December 1970 storm.

of frequencies.

The results are therefore consistent with the spectral gap, at least in the range between periods of 10 minutes to one hour, in so far as no significant peaks were found in this range. Its existence cannot, however, be proven from the results since nothing can be said about the spectra on either side of the frequency range.

11. Precipitation

According to classic foehn theory, moisture is condensed and removed from air which is lifted up the windward slopes; the latent heat liberated during this process will be realized as warming at the leeward stations. Ives (1950) used this principle to define his cases in a study of Chinook winds in Colorado. In order to be classified as 'meterological' Chinooks (as opposed to other warm winds) according to Ives, westerly winds at Denver had to be accompanied by precipitation at a windward station (Fraser) and potential temperature at Denver had to be higher than at Fraser. Cook and Topil (1952), on the other hand, using the more usual definition of warm and dry descending wind, did not find a precipitation shield on the windward side to be necessary for Chinooks east of the mountains in Colorado. Observing that the air between 840 and 700 mb over Denver was 7°C warmer than over Grand Junction, they estimated that this would require about 4 mm of precipitation, but found that only a few stations west of the Divide had reported showers, none of which heavy.

To determine the importance of windward precipitation for the 20 Boulder wind storm cases, the published 'Hourly Precipitation Data' (U.S. Department of Commerce) were examined for the nearest two west slope stations, Grand Lake and Hot Sulphur Springs (about 50-70 km west of Boulder). Assuming an average west wind component at 700 mb of about 15-20 m s^{-1} (based on upper air data presented in Section 13), a two hour time lag, at most, would compensate for the time it took an air parcel to cross the mountains. The definition of the onset of wind storms, in terms of a beginning of hourly west wind components ≥ 10 m s^{-1} at the Boulder stations, omitted short-lived wind speed peaks associated with frontal passages. This eliminated the problem of precipitation resulting from the action of a front preceding the wind storm. Thus, starting two hours prior to the onset of the storms, the hourly precipitation at the two stations was noted for the entire wind storm period. The results in Table 11.1 show that out of the 20 storms there was some precipitation in only 5 cases (cold, warm and indifferent storms), generally amounting to about 0.25 - 0.50 mm hr^{-1}. In three cases it lasted only through part of the storm and in only one did both stations record precipitation. The latter case, the 11 February 1971 storm (an indifferent wind storm), was complicated by the fact that a front immediately followed the storm and it is therefore not obvious whether the air warmed by the release of latent heat was identical with the air involved in the wind storm at the foot of the eastern slopes.

Table 11.1. Hourly precipitation data during Boulder wind storms with precipitation on the western slopes.

Storm date	Hours	Temperature characteristic of the storm	Grand Lake day : hour / amount (mm)		Hot Sulphur Springs day : hour / amount (mm)		
7 April 1969 to 8 April 1969	1000 0900	cold			7:1300 0.25	7:1700 0.25	7:1900 0.25
25 January 1970 to 26 January 1970	0200 0800	cold	25:0600 0.25	25:0700 0.50			
3 February 1970 to 4 February 1970	0600 0400	warm	3:2300 0.50 4:0100 0.25	3:2400 0.50 4:0200 0.75			
20 January 1971	1700- 2300	cold	20:1800-2400 total: 2.75				
11 February 1971	0100- 0300	indifferent	10:2300 0.50 11:0100 0.75	10:2400 0.25 11:0200 1.00		11:0100 0.75	11:0200 1.00

The warming effect of 0.25 mm of precipitation can be calculated from the relationship

$$\partial T = \frac{L_v \; dm_v}{m_a \; c_p}$$

where T = temperature
L_v = latent heat of vaporization
m_v = precipitation
m_a = mass of air
c_p = specific heat at pressure.

Assuming an average density of 0.9 kg m^{-3}, a column of air between about 820-650 mb (both figures are based on data presented in Section 13) would be heated by about 0.3°C. This means that even for the 11 February 1971 case, with a precipitation value 4 times the above, the amount of heating was still insignificant (i.e., 1.2°C, assuming that the air warmed by the release of latent heat was actually involved in the Boulder wind storm).

It is therefore concluded that <u>although some precipitation was recorded during 5 of the 20 cases, the amount was insufficient for the classic 'foehn effect' in any of the cases</u>. Consequently, <u>the storms identified as warm in Section 8 must have been entirely the result of warm air advection and subsidence</u>.

12. Surface Pressure

Surface pressure gradients forcing air flow across mountains on a synoptic scale have long been assumed to be part of the mechanism of downslope winds around the world (see Section 2).

In addition, local pressure minima on the lee side have been found to be associated with wind storms and are considered to be a consequence of the downslope flow. Lilly and Zipser (1972) have, for instance, shown that during a Boulder storm local pressure dips of 5-7 mb occurred between the Divide and Denver (Fig. 12.1).

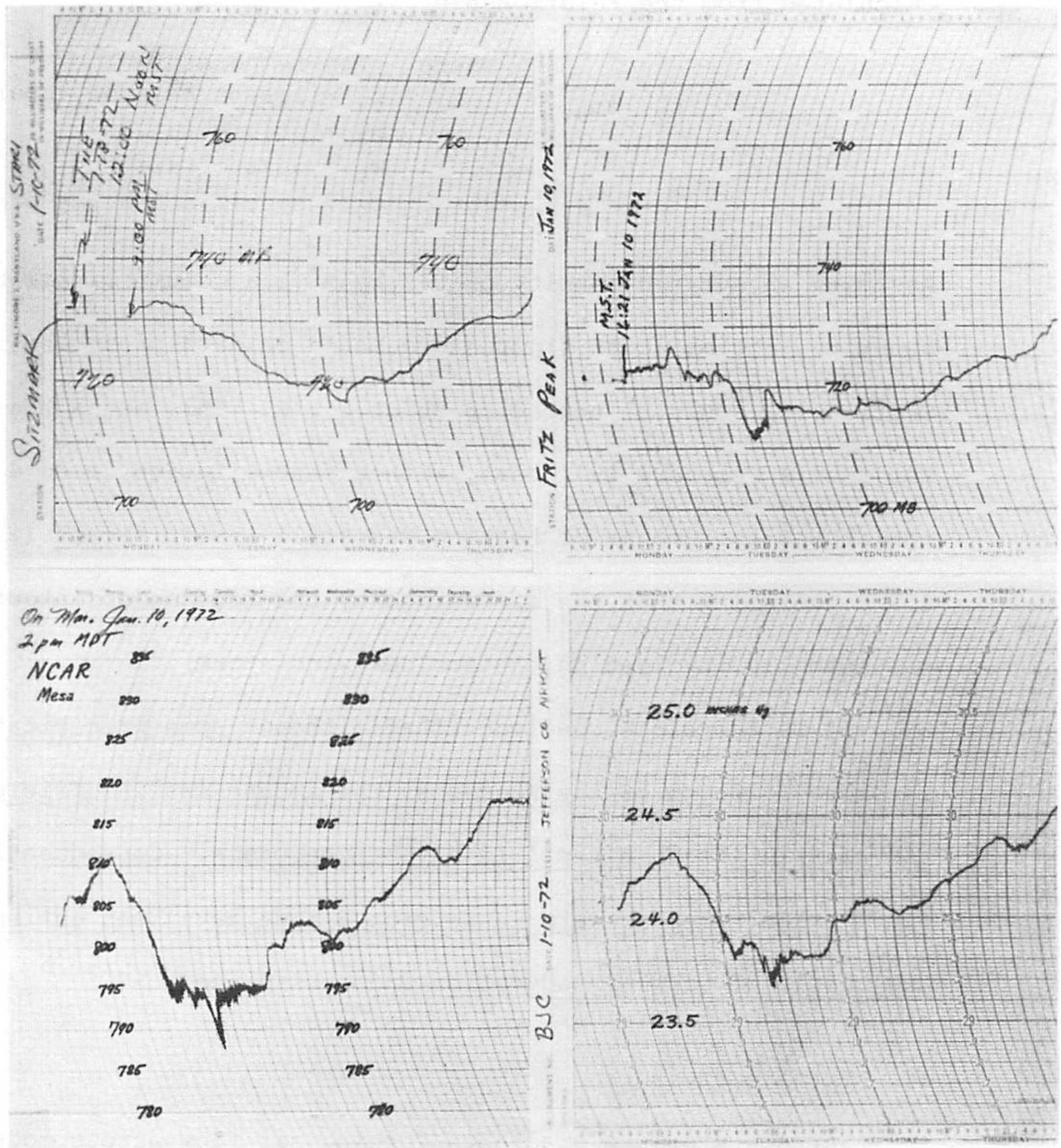


Fig. 12.1. Four microbarograph traces from the Front Range area on 11 January 1972. Upper left: western side of Continental Divide at about 9000 ft. Upper right: eastern side of Divide at about 9000 ft. Lower left: NCAR site at 6100 ft. Lower right: Jefferson County Airport at 5500 ft. All records are in millibars except for the last, which is in inches of Hg. Note asterisks placed on chart to indicate noon and midnight on the 11th (from Lilly and Zipser 1972).

These appeared to be directly associated with (i.e., occurred below) the extremely warm region in the deep trough of the 'lee air flow disturbance' over the eastern slopes (Fig. 12.2), which is hydrostatically consistent.

Additional interest in the surface pressure gradient across the mountains during wind storms comes from the fact that the torque produced by topography on the atmosphere during a single severe storm in Boulder can be comparable to the global total for this latitude and season and thus have a significant effect on the general circulation (Lilly 1972). The mean torque imparted by all topographic features around the globe in the belt 40-45°N in winter is estimated to be about -1.3×10^{25} gm cm^2 s^{-2} (Newton 1971).

For the wind storm cases considered in this study, hourly station pressure was obtained for the lee stations Denver and Boulder. Data for the windward station, Eagle (about 135 km west of Boulder), were not available for the year 1971 and the following computations therefore cover only 13 of the 20 wind storm cases.

Mean deviations from normal station pressure (according to U.S. Standard Atmosphere) were calculated for the hour with the highest west wind component at the Boulder stations and for an hour 'prior' to the wind storms. The latter was determined by inspecting the pressure trend before and during the storm periods and noting the earliest hour this trend began at any of these stations. In a few cases the hourly data did not cover a sufficiently long period to include the beginning of this trend and the first hour of the period covered was used.

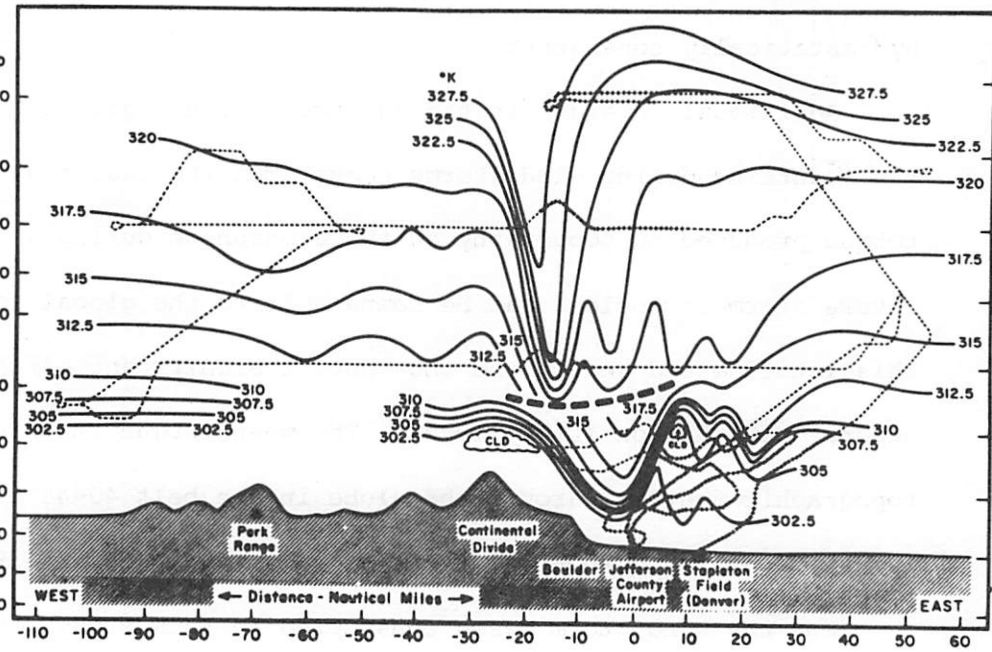


Fig. 12.2. Cross section of the potential temperature field in degrees K over the mountains and foothills as obtained from analysis of the Queen Air and Sabreliner data on 11 January 1972. Data above 500 mb are exclusively Sabreliner taken from 1700-2000. Data below 500 mb are primarily Queen Air taken from 1330-1500 MST. Flight tracks are indicated by the dashed lines, except by crosses in turbulent portions. At this time it is not possible to determine whether the apparent westward displacement with height of the major features is real or related to the time difference between the two flights. Windstorm conditions on the ground extend eastward to the location where the isentropes rise sharply, a few miles east of the origin at the Jefferson County Airport (from Lilly and Zipser 1972).

Fig. 12.3 shows the mean pressure deviations across the Rocky Mountains for the times 'prior' to and during wind storms in Boulder. (For comparison purposes, the mean west wind component at the lee stations for the same priods are also shown.) In the mean, pressure on the windward slope 'prior' to wind storms in Boulder was nearly normal while in the lee the pressure was already somewhat low. The mean time difference between 'prior' to and the beginning of the storm (first hour with west wind components at any Boulder station ≥ 10 m s^{-1}) was at least 8 hours (considering the fact that the beginning of the trend was not always covered by the data), and the time difference between 'prior' and during wind storms (the hour of maximum wind in Boulder) was at least 12 hours. Over the latter period of 12 hours, the windward pressure increased slightly by about 2 mb, to 2 mb above normal while at the lee stations it dropped further by about 1 mb, to 6 mb below normal at Denver and by about 4 mb, to 11 mb below normal at Boulder.

A closer look at the pressure trends before and during wind storms in Boulder shows that the surface pressure gradient across the mountains may be due to different processes, although the end result may be the same. As will be shown below, the trend is not always a decrease, as usually assumed, but quite often an increase. In general, a pressure rise was found to first begin at the windward station whereas a pressure decrease was usually first noticeable at the lee stations, the mean difference between the beginning of this trend at the windward and lee stations being about two hours.

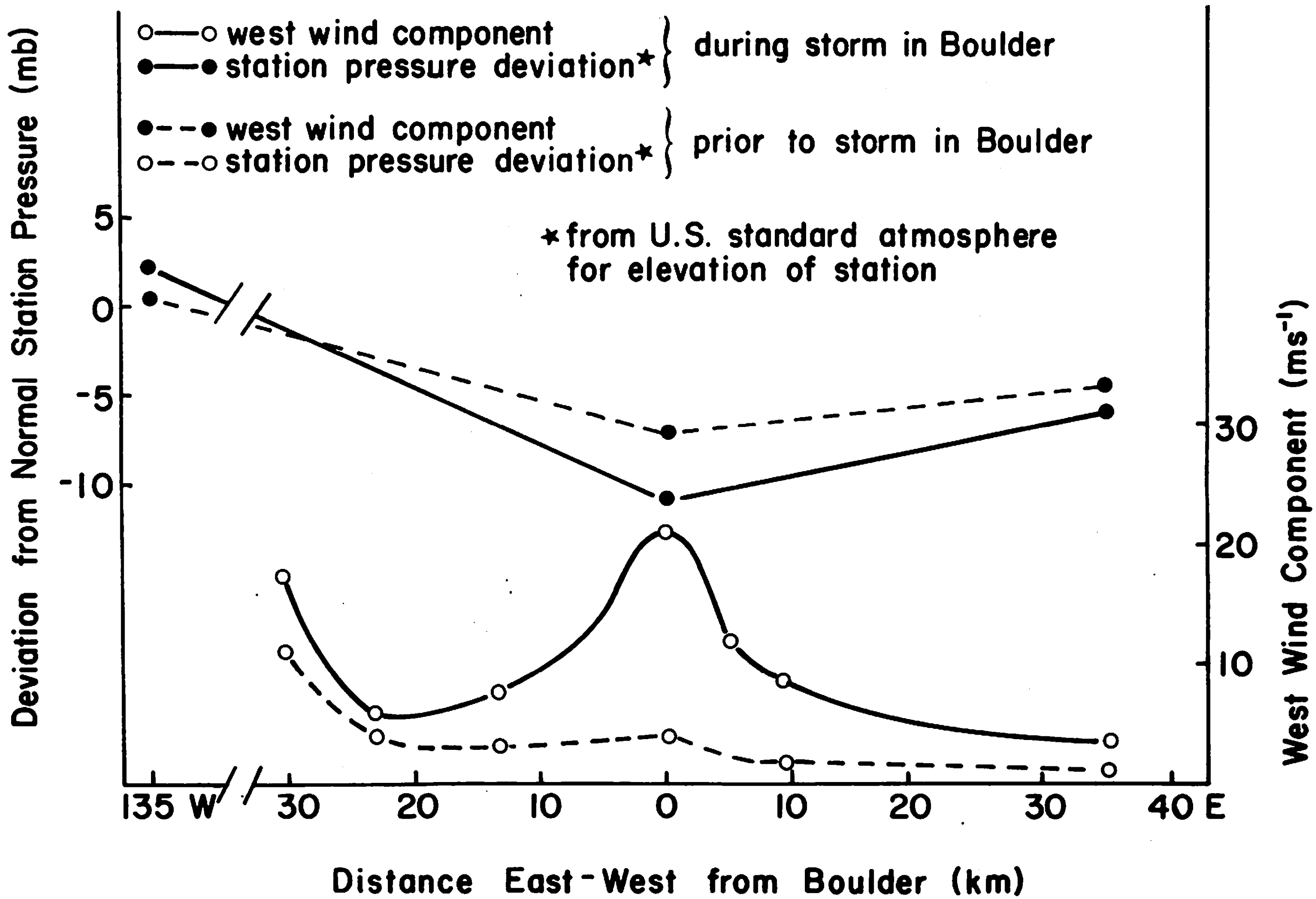


Fig. 12.3. Cross section of mean deviations from normal station pressure and mean west wind components for times prior to and during wind storms in Boulder.

The average difference of 12 hours between the beginning of the pressure trend and the height of the wind storm is used in Fig. 12.4 which presents the results of Fig. 12.3 in a different form emphasizing change over time. Fig. 12.4 A shows the means for all 13 cases, Figs. 12.4 B, C and D give the means for the cold, warm and indifferent cases (as defined in Section 8). The difference between the cold and warm storms is outstanding. <u>For the warm cases the pressure change at all three stations was negative</u>, decreasing by about 4 mb at Eagle, 7 mb at Denver and 11 mb at Boulder. <u>For the cold storms, on the other hand, the pressure change at all stations was positive</u>, increasing by about 2 mb at Boulder, 3 mb at Denver and 7 mb at Eagle (the difference in trends between the cold and warm cases is statistically significant). However, regardless of the type of storm, the pressure tended to be above normal on the windward side and below normal in the lee at the height of the storm.

The physical explanation of the difference in pressure trends lies in the difference in synoptic situations leading to the cold and warm wind storms in Boulder. The cold storms, belonging to the class of bora-type winds, involve a cold anticyclone moving into the region and/or intensifying west of the mountains. Consequently, there is a pressure increase at all stations, but the increase begins first at the windward station and is twice as rapid there as the pressure rise at the lee stations. At the height of the storm the pressure is, therefore, above normal at the windward stations but still below normal in the lee. In the case of the warm winds, on the other hand, a low pressure system is the

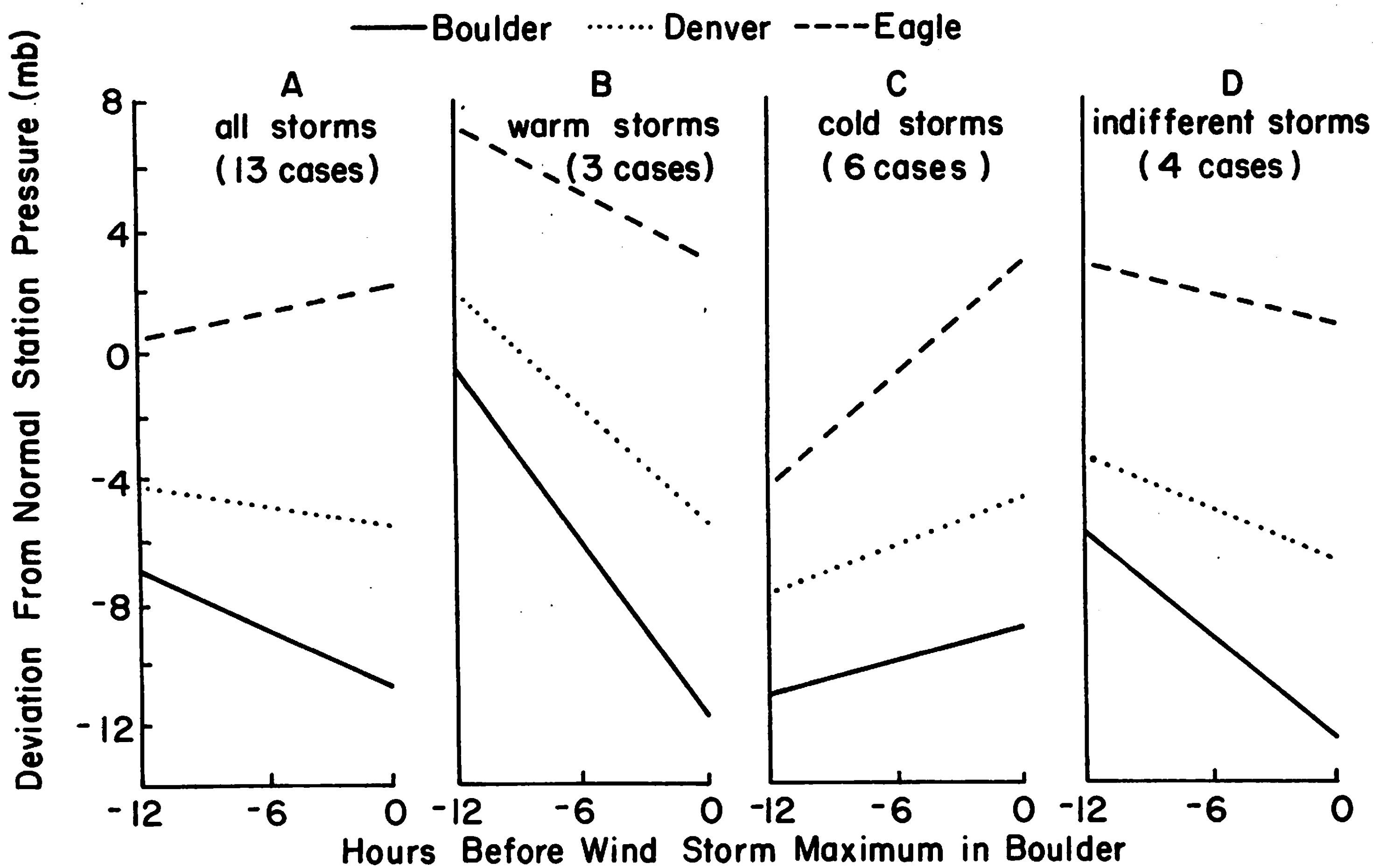


Fig. 12.4. Changes in mean deviations from normal station pressure between 12 hours prior to and the height of wind storms in Boulder at stations on both sides of the mountains, classified by storm type.

more dominant feature, moving into the region and/or deepening east of the Rocky Mountains. Consequently, there is a pressure decrease at all stations, but the decrease begins first and is twice as rapid in the lee as at the windward stations. At the height of the storm the pressure is, therefore, below normal in the lee but still above normal at the windward stations. In both cases, <u>the end result of the pressure change is a total difference from normal station pressure between Eagle and Denver of about 8 mb which is the same for both, the cold and warm wind storms</u>.

Pressure trends for the indifferent storms are very similar to those for the warm cases (the difference between the two is statistically not significant). This suggests that many of the indifferent cases could have been warm storms but were not classified as such because the temperature change took place long before the onset of the storm in Boulder (i.e., more than 6 hours).

The difference between mean deviations from normal station pressure at Denver and Eagle during wind storms in Boulder, 8 mb, may be considered to represent a mean synoptic scale pressure gradient across the mountains of 4.7 mb/km. The torque upon the atmosphere resulting from this pressure difference of 8 mb between the east and west faces of the mountains may be estimated from

$$T = r \cos \phi \int_{0}^{H} P \, dz$$

where
T = torque per unit distance
r = earth's radius
ϕ = latitude
H = height of mountain range
P = pressure.

Assuming a linear decrease in the horizontal pressure gradient with height to zero at mountain top levels over a vertical distance of 2 km, the resulting torque is -4.24×10^{24} gm cm^2 s^{-2} over a distance of 1° latitude. The mean torque produced by the mountains on the atmosphere during the 13 wind storm cases was therefore almost twice that for the entire globe at this latitude and season. The effect on the general circulation, in terms of a decrease in the momentum of the atmosphere and a consequent slowing down of the westerlies, must be substantial.

The difference of 5 mb between Boulder and Denver is the mean local pressure minimum, associated with the wind storms, toward which wind speeds would be accelerated from the west. At the same time, it represents an adverse pressure gradient to the east between Boulder and Denver. The change in wind speed corresponding to such a pressure change may be estimated from Bernoulli's equation

$$\partial \left(\frac{v^2}{2} + \frac{p}{\rho} + gh \right) = 0$$

where v = wind speed
p = pressure
ρ = density
g = gravitational acceleration
h = height

which is an expression of the conservation of energy of an air parcel along a streamline.

As a first approximation it is assumed that the height of the streamline above ground is constant. Taking mean gust velocities to represent wind speeds of the air above the surface boundary layer, a local pressure dip of 5 mb would accelerate the air from 21 m s^{-1} at Gold Hill to 38 m s^{-1}. During wind storms in Boulder, the mean gust at Southern Hills is 33 m s^{-1} (taking the height factor for the station into consideration). The difference between the estimated and observed values could be explained in terms of frictional dissipation. The local pressure dip is therefore sufficient to account for the increase in wind speed down the slopes. For zero velocity at Gold Hill, the same local pressure minimum would accelerate the air to 32 m s^{-1}. This explains why no relationship between wind on the slopes and wind in Boulder can be found and velocities in the mountains may actually be very low during some of Boulder's wind storms.

Using the same assumptions, the pressure increase of 5 mb between Boulder and Denver would correspond to a decrease in velocity from 33 m s^{-1} at Southern Hills to 11 m s^{-1}. No Denver gust data were available, but for a mean hourly speed of 4 m s^{-1} at Denver during storms in Boulder (taking the height factor for the station into consideration) a mean gust factor of 2.8 may be assumed from the curves shown in Fig. 10.7 (Section 10). This factor would give a mean gust of 11 m s^{-1} at Denver. Thus, assuming mean gusts to represent wind speeds above the surface boundary layer, the calculations suggest that the adverse pressure gradient is sufficient to account for the observed decrease in kinetic energy. Fig. 9.23 (Section 9) shows that during storms in Boulder the mean gusts decrease rapidly with increasing distance to the east. Extrapolation of this decrease suggests that mean gusts of 11 m s^{-1} may already be found about 15 km east of the Southern Hills station. Since this is also the distance over which most of the mean wind speed decrease takes place, one might conclude that most of the 5 mb pressure gradient is also concentrated over this short distance. This would be in agreement with observed almost vertical updraft regimes above Boulder as shown, for instance, in Fig. 12.2.

13. Air Stream Characteristics

Knowledge of temperature and wind characteristics of the air mass crossing the mountains during Boulder wind storms is almost as fragmentary as that of temperature and wind distributions at the surface. No systematic analysis has ever been done, only the occasional case study of an extreme event such as the 7 January 1969 storm.

Denver temperature profiles during a Chinook wind case in eastern Colorado have been discussed by Cook and Topil (1952). These soundings show some tendency toward stability at mountain top levels (between about 600 mb and 700 mb). However, a more detailed review of their results was not considered to be very useful for the purpose of the present study because the definition used was not in terms of wind speeds. The same applies to Beran's (1967) analysis of Chinook wind cases.

Harrison (1956) has studied the synoptic characteristics of the atmosphere during periods of strong mountain wave activity in the Denver area, defined in terms of severe clear air turbulence. The mean wind speed profile at Denver for these lee wave periods (Fig. 13.1) shows a minor wind maximum of 23 m s^{-1} at about 600 mb (slightly above mountain top levels), an average shear of about 6 m s^{-1} km^{-1} below it and a somewhat weaker shear above. Associated upwind temperature profiles at Grand Junction (Fig. 13.2) indicate an inversion or stable layer with the base varying between 780 and 610 mb. In the mean temperature profile (Fig. 13.3), however, this stable layer is hardly detectable, because straigth

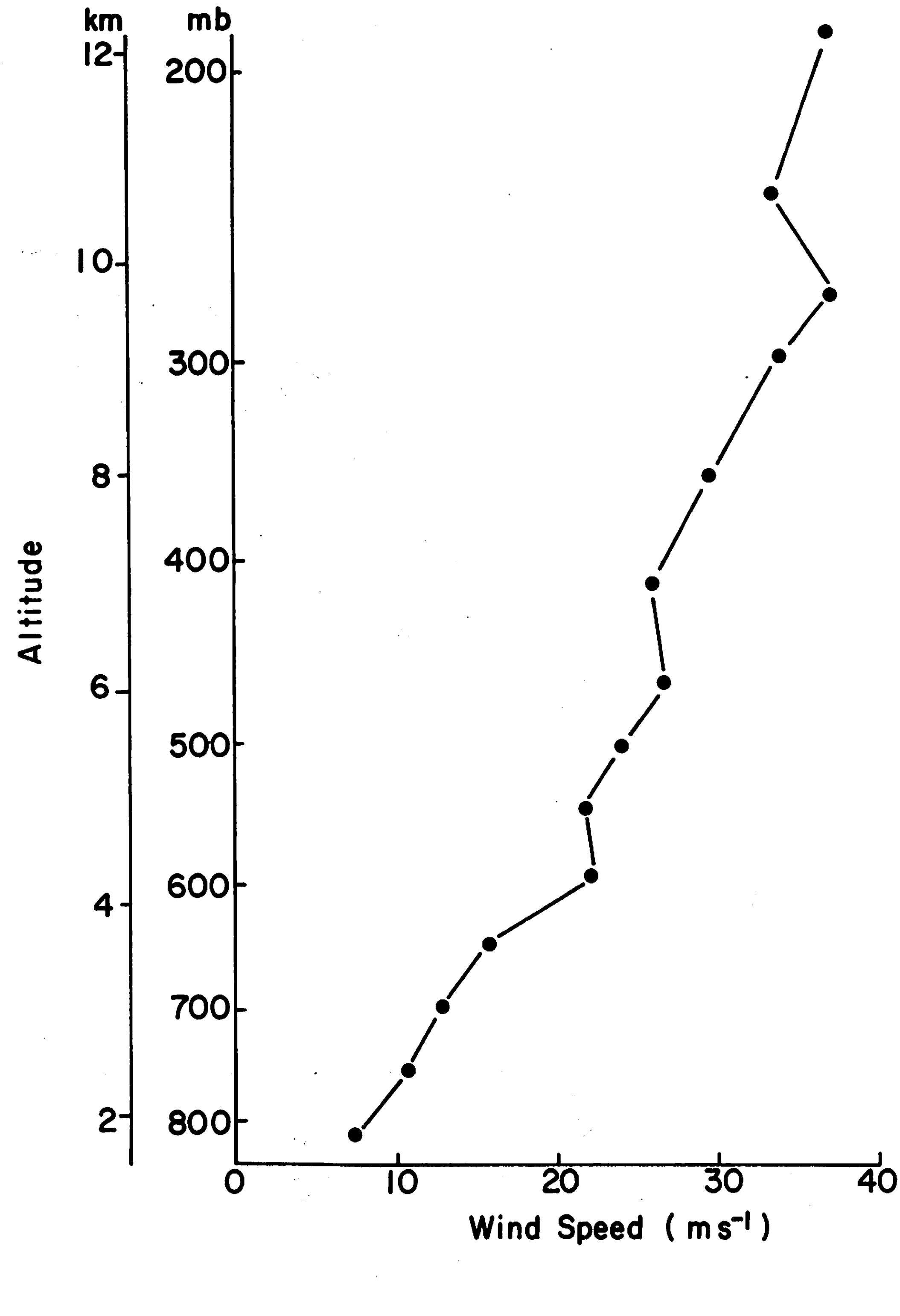


Fig. 13.1. Mean wind speed profile at Denver for 7 strong mountain wave cases (after Harrison 1956).

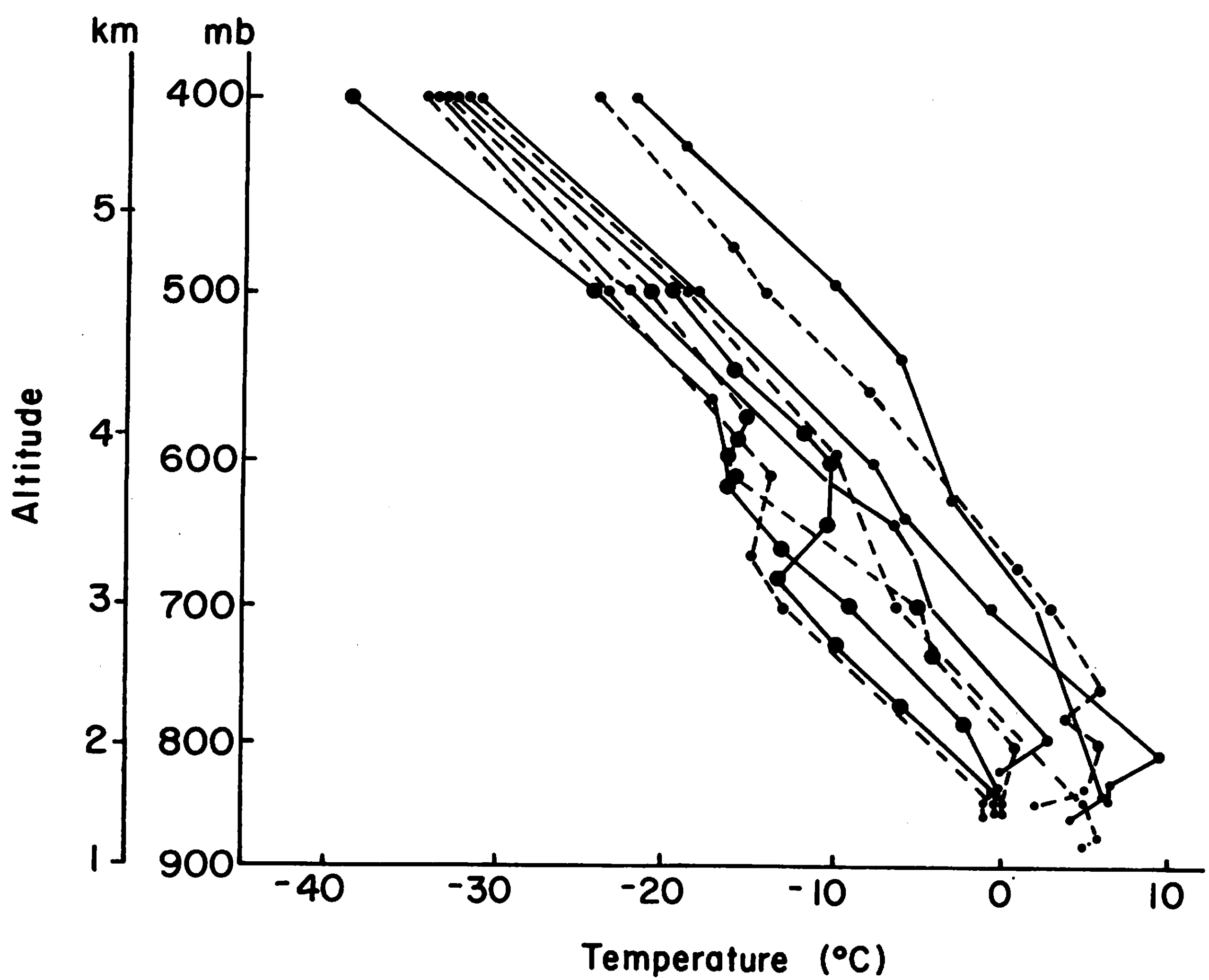


Fig. 13.2. Temperature profiles at Grand Junction for 9 strong mountain wave cases, strongest cases emphasized by large dots (after Harrison 1956).

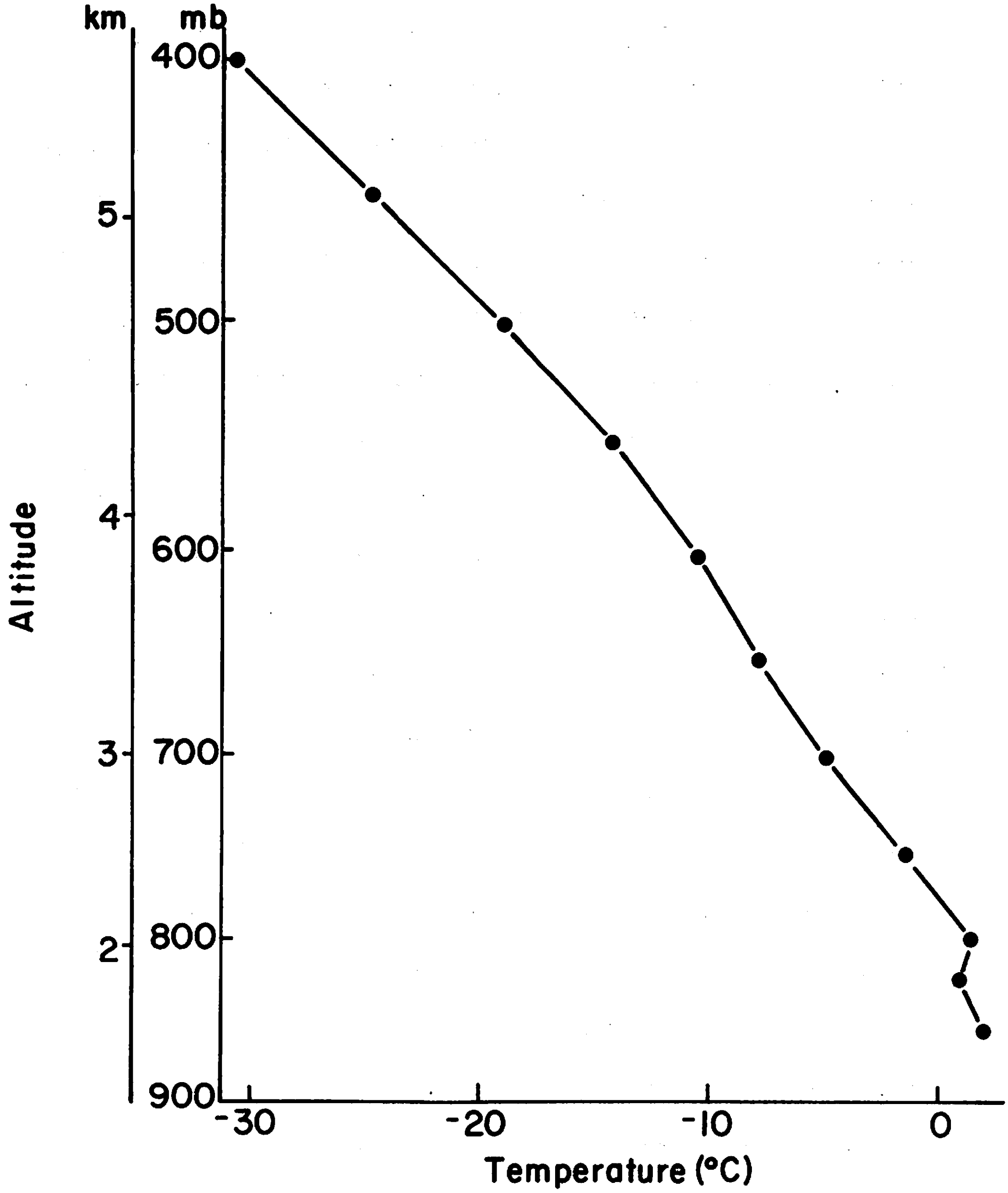


Fig. 13.3. Mean temperature profile at Grand Junction for 9 strong mountain wave cases (after Harrison 1957).

averaging resulted in excessive smoothing. The stable layer around 800 mb represents a strong surface inversion.

Danielsen and Bleck (1970) and Vergeiner (1971) used as input into their linear models detailed data gathered during observational lee wave programs at Boulder as well as Denver radiosonde data for the time of onset of a few Boulder wind storms. They concluded that the difference between lee waves and storms may lie in the strength of the wind at mountain top levels and the shear below. Houghton and Isaacson (1968) used as input into their non-linear hydraulic jump model data collected during a lee wave program on a day with some wind on the slopes west of Boulder and found that a lowering of the initial inversion level by 300 m (equivalent to about 20 mb at that level) caused the leeward location of the surface wind maximum to move from close to mountain top to halfway down the slopes (Fig. 3.4). Results from models currently being developed by Klemp and Lilly (1973) also show sensitivities to changes in air mass characteristics.

A major handicap in the study of Boulder wind storms has been the lack of adequate upper air data. Because of the, so far, somewhat unpredictable nature of the phenomenon and its short duration, it has been rather difficult to have an aircraft ready to fly into intense storms (Lilly and Zipser 1972). Furthermore, the fact that they tend to either begin or end at about the time of sunrise or sunset (see Section 7) and, therefore, usually occur between times of radiosonde ascents in this part of the country has also been inconvenient.

Another problem lies in the necessity to use data from a station downwind of the disturbance to estimate properties of the undisturbed windward air, since the disturbance may distort the air flow characteristics. For instance, air stream deformation and turbulence will have an effect on wind shear and lapse rate. The sharpness and location of an inversion recorded by a radiosonde sounding in a zone of wave development may differ from that in the undisturbed stream (Scorer 1953). Also, hydraulic jump models suggest that the level of the inversion or stable layer downwind of the jump may be lower than its initial position (Kuettner 1959; Houghton and Issacson 1968).

It is therefore of interest to determine the mean upwind as well as downwind air flow characteristics during the 20 wind storm cases and to compare the two.

a. Upwind soundings. Determination of the characteristics of the undisturbed air flow for wind storms in Boulder is somewhat difficult since there is no radiosonde station on the windward slopes. Choice of the upwind station should therefore be dependent upon the direction of the approaching air stream; but because Grand Junction is closest (about 300 km and 15° S of W of Denver) its soundings have frequently been assumed to represent upwind conditions (e.g., Harrison 1956; Sangster 1970). However, wind direction at Denver in the layer between the surface and at least 350 mb at the time of storms in Boulder is on the average north of west (a point to be discussed further). A better upwind station might therefore be Salt Lake (15° N of W but also 600 km and another mountain range upwind of Denver), and in some cases even Lander

(55° N of W and 450 km upwind). Determination of the appropriate upwind station is however difficult because of the spacing of these stations as well as the 12-hour interval between soundings. The problem could possibly be overcome by computing three-dimensional trajectories and interpolating with respect to time and space, but for the purpose of this study a simpler approach was considered sufficient and superior to the assumption that Grand Junction was always the upwind station.

As a first approximation, the 500 and 700 mb constant pressure maps for the soundings about the time of the wind storm in Boulder (one or more, depending on the duration of the storm) and the sounding 12 hours prior were examined. Possible upwind station(s) were determined on the basis of wind directions at the three stations, Grand Junction, Salt Lake and Lander, at the two pressure levels and their changes between sounding times. Also considered was the approximate time it would take an air parcel to travel between the upwind station and Denver. This frequently resulted in a choice between two stations and/or sounding times. The temperature profiles for these stations and times were then examined for the presence of the 'storm air', using as a guide the potential temperature of this air as it appeared in the Denver sounding. In some cases the final choice was rather arbitrary and subjective.

For each of the 24 upwind soundings (one per storm plus an additional sounding for 4 of the more persistent ones), the upper limit of the storm air as indicated by a marked change in lapse rate, i.e., the base of a more stable layer, was noted. Potential temperature as well as wind speed and direction at 50 mb intervals were obtained from the routinely published upper air observations; in addition, potential temperature for the level 25 mb above and below the base of the stable layer were interpolated. In order to overcome the problem of excessive smoothing resulting from straight averaging, a composite sounding was produced by displacing the values up and down with respect to the base of this layer. A representative level of the base was chosen by grouping pressure values at the base into 25 mb intervals and determining the central value of the modal class. This method was used to reduce the effect of a few extreme values, which was perhaps more important for the Denver soundings (discussed below) in which downwind effects presented more of a problem. The modal location of the base of the stable layer was found to be 575 mb, which is in agreement with Harrison's results.

Outstanding features of the mean upwind sounding (Fig. 13.4) are the approximately 25-mb-deep isothermal layer and the steady increase in wind speed from the surface up. Across and below the stable layer there is a strong vertical wind shear of about 6.4 m s^{-1} km^{-1} (about twice the mean for this area and season (Ratner 1959)), a mean west wind component of 18.6 m s^{-1} at mountain top levels and 24.1 m s^{-1} just above the stable layer. At upper levels the shear is about 3.6 m s^{-1} km^{-1} with the mean speed increasing to a maximum of 40 m s^{-1} at about 300 mb, which also appears to be the mean level of the tropopause.

Separation into warm, cold and indifferent storms, as defined in Section 8, resulted in the three profiles shown in Fig. 13.5

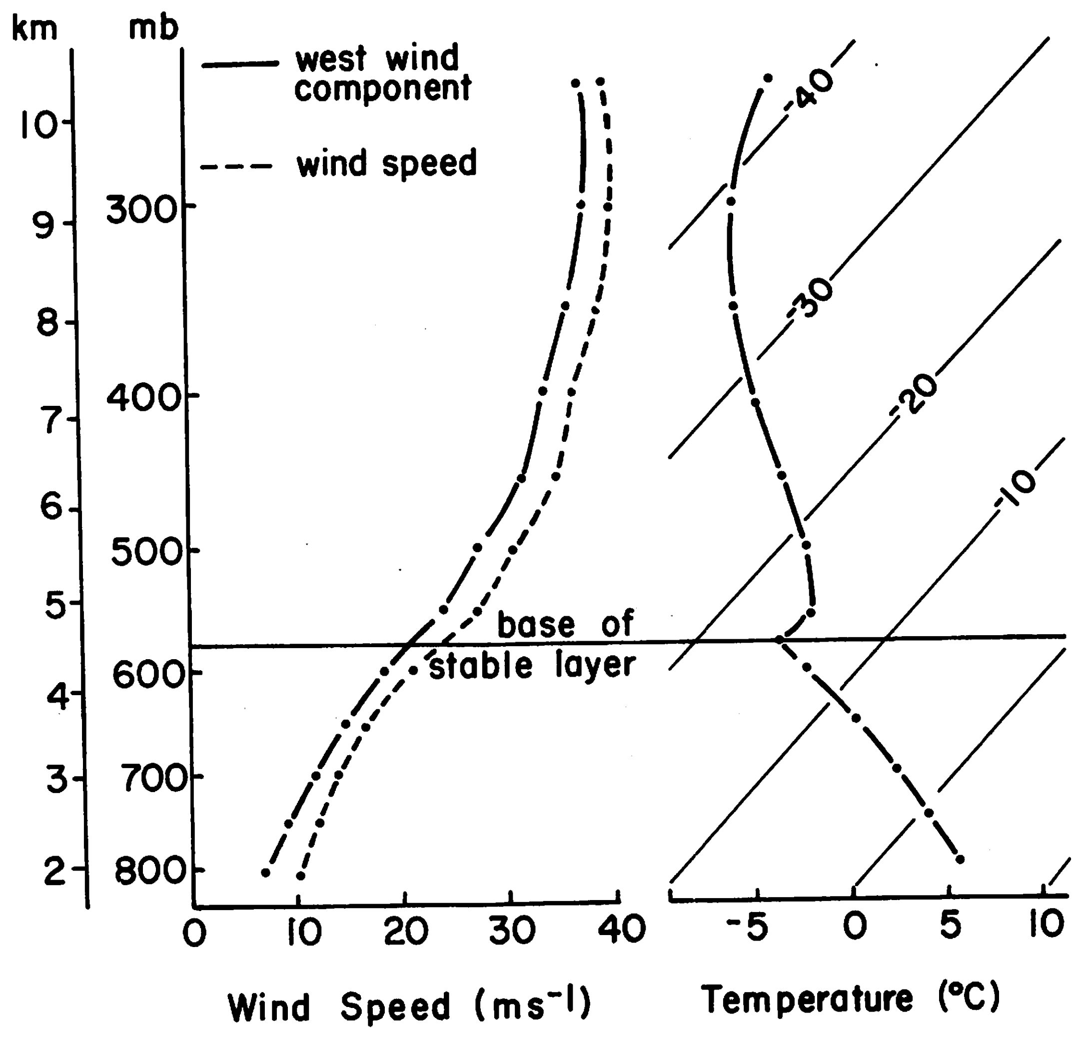


Fig. 13.4. Composite upwind sounding for time close to the beginning of wind storms in Boulder.

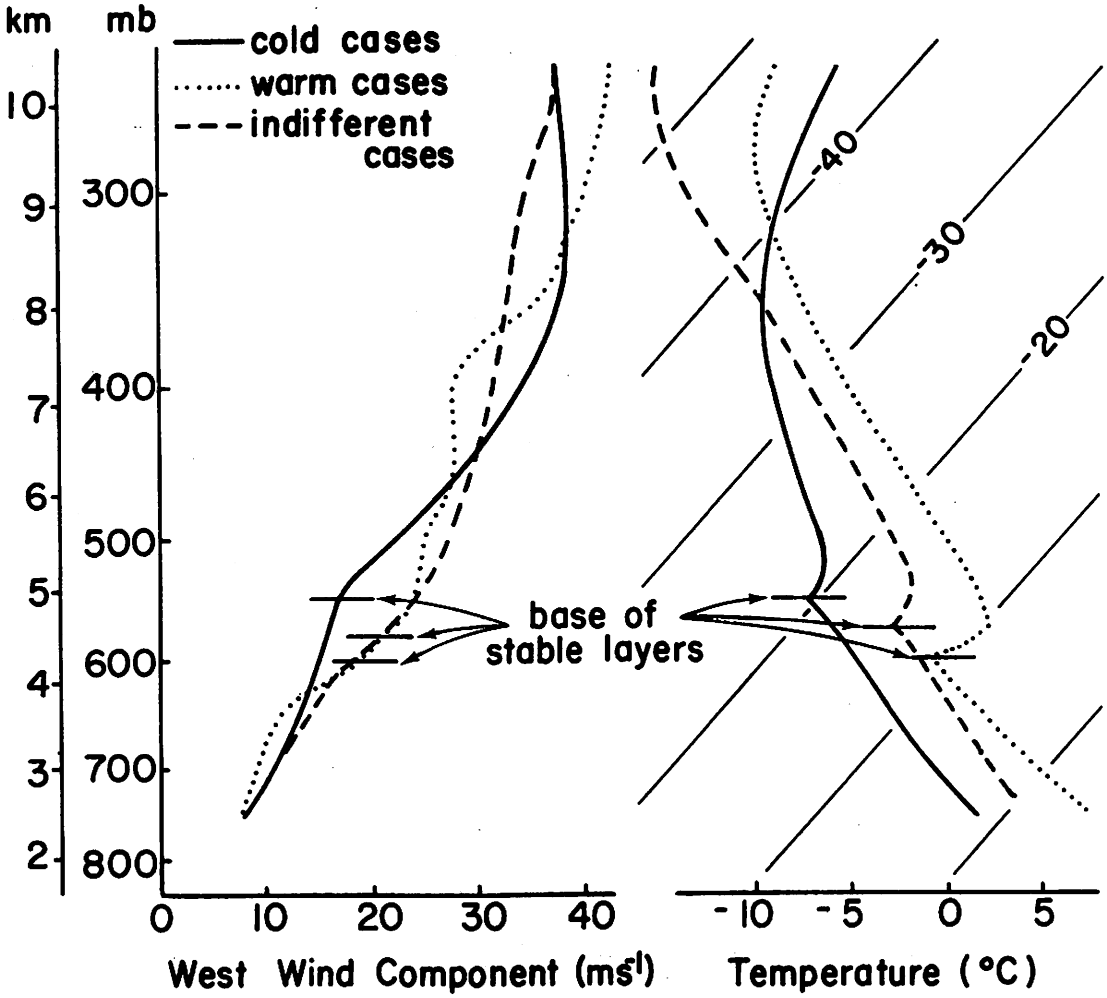


Fig. 13.5. Composite upwind sounding for time close to beginning of cold, warm and indifferent wind storms in Boulder.

(made up of 5, 10 and 9 soundings respectively). The location of the stable layer was found to be slightly different for the three types, i.e., at 550 mb for the cold cases, at 575 mb for the indifferent ones and 600 mb for the warm storms (these differences are statistically not significant); the stability of the layer for the warm cases is however significantly greater. Below this layer, there is a significant temperature difference of 3°K between the cold and warm storms, while the apparently stronger wind shear for the warm cases is not statistically significant. Above the stable layer, the mean west wind component is about the same for cold and warm cases, i.e., 20 m s^{-1}; but because of the difference in its location the wind speeds between mountain top levels and 500 mb are lower for the cold cases (the maximum difference of about 6 m s^{-1} at 550 mb is however non-significant statistically). The cold storms also have a higher stability above the stable layer and a lower tropopause. The indifferent storms seem to be somewhat similar to the warm ones.

For comparative purposes, mean wind speed profiles at Grand Junction for the winter months (December) 1947-1952 are shown in Fig. 13.6 together with an indication of the variability (Salt Lake data were not available). For the same months the 35-55°N zonal index at 700 mb for the western hemisphere was about 11 m s^{-1} which is the same as the long term mean for the period 1944-1963. The 1947-1952 reference speeds in Fig. 13.6 may therefore be considered to be representative of long term average conditions, at least at lower levels. The index for the December months

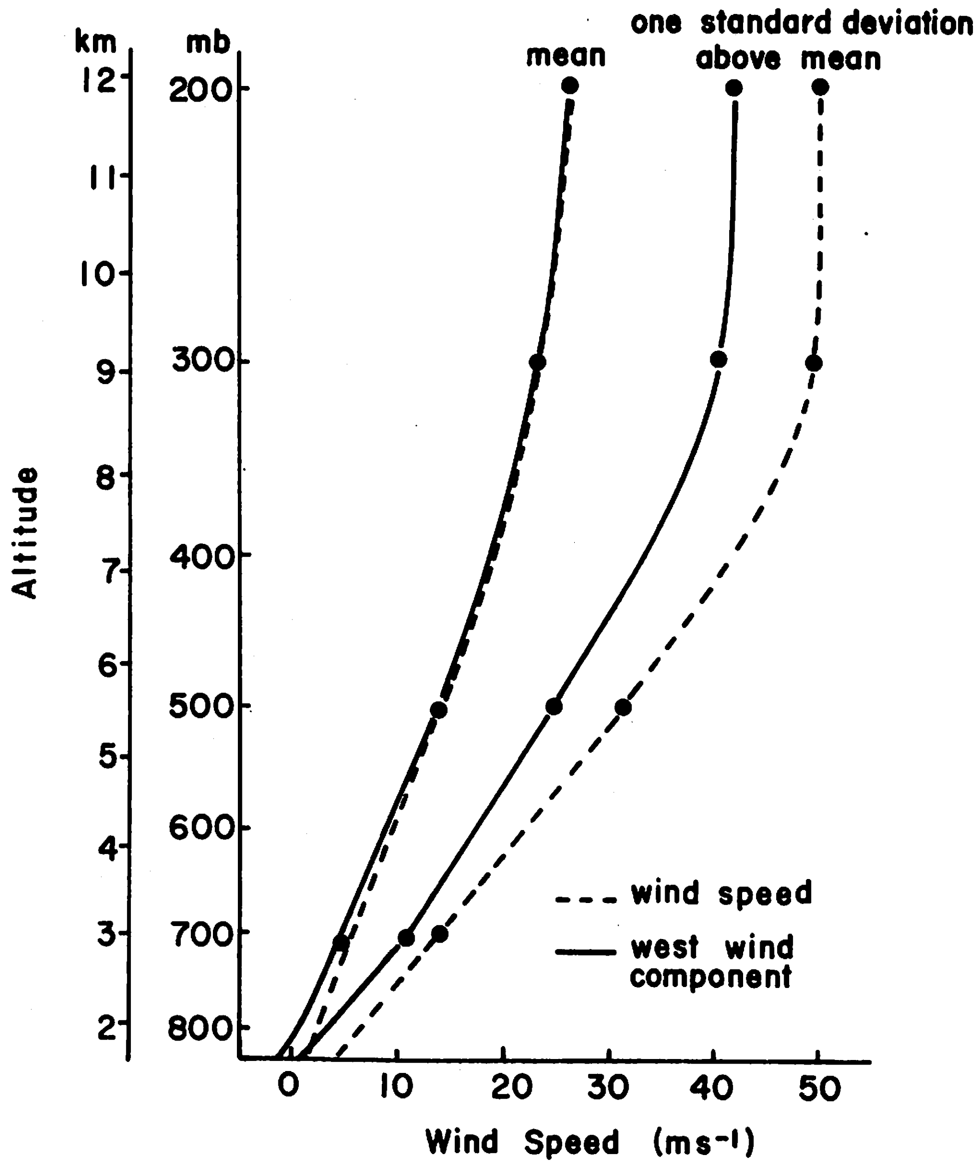


Fig. 13.6. Mean wind speed profile at Grand Junction, winter (December) 1947-1952, data from Crutcher (1958).

1968-1970, representative of the period covered in this study, was about 10 m s^{-1}.

From Fig. 13.6, upstream winds in the lower troposphere to just above the stable layer during wind storms in Boulder were about one standard deviation (or a factor of two) above the 5-year mean for Grand Junction and were less than that in the upper troposphere, above 450 mb. Furthermore, mean gusts at the surface during wind storms in Boulder (i.e., 36 m s^{-1}) were much higher than the mean wind up to the 400 mb level in the undisturbed flow. Riehl (1971) commented on the 'unusually weak winds' at upper levels (around 250 mb) during the 30 November 1970 wind storm (one of the most severe cold storms considered in the present study, see Appendix A); but <u>mean upper tropospheric winds were found not to be excessively high in the upwind sounding during the 20 wind storm cases (generally less than 40 m s^{-1})</u>.

b. Downwind Soundings And Comparison With Upwind Temperature And Wind Profiles. To determine the downwind air flow characteristics of the 20 wind storm cases, all Denver soundings for times close to or during these storm periods were studied. Knowing the slope temperature during each of the storms and taking into account the tendency for the free air to be about 4°K warmer than the slope air (see Section 8), these soundings were examined for the presence of the storm air. For each of the 32 soundings in which this air was found (almost two per storm), its upper limit as indicated by a marked change in lapse rate was noted and a representative height was again chosen by determining the model class; this was found to be around 650 mb. Potential temperature above and below this level was abstracted at 20 mb intervals from plotted temperature profiles and wind data at 50 mb intervals were obtained from the routinely published upper air reports. A composite sounding (Fig. 13.7) was again produced by using the base of the stable layer as a reference point and averaging the temperature and wind data with respect to it.

Comparison between the mean upwind and downwind soundings shows close agreement (within one degree) in the potential temperature of the air at the level of the stable layer and through a 100-mb-deep layer below it. The two air masses can therefore be assumed to be indentical and the upwind soundings, although determined by relatively simple methods, may be taken to be representative of the undisturbed flow (at least at lower levels).

The downwind temperature profile shows a 20 to 25-mb-deep isothermal layer marking the upper limit of the storm air, which is little different from the 25-mb-deep layer in the upwind profile. The base of this layer downwind, at about 650 mb and thus below mountain top levels, is considerably lower than the upwind base above mountain top levels, at 575 mb (the difference of 75 mb is statistically significant just above the 5 per cent level). This difference may result from the stable layer having a sloping surface. But the fact that the downwind base of this layer, which represents the upper limit of the storm air, was somewhat below summit levels suggests that at least part of the differences between the upstream and downstream levels was caused by the disturbance. Figs. 3.2 and 9.22 would support this.

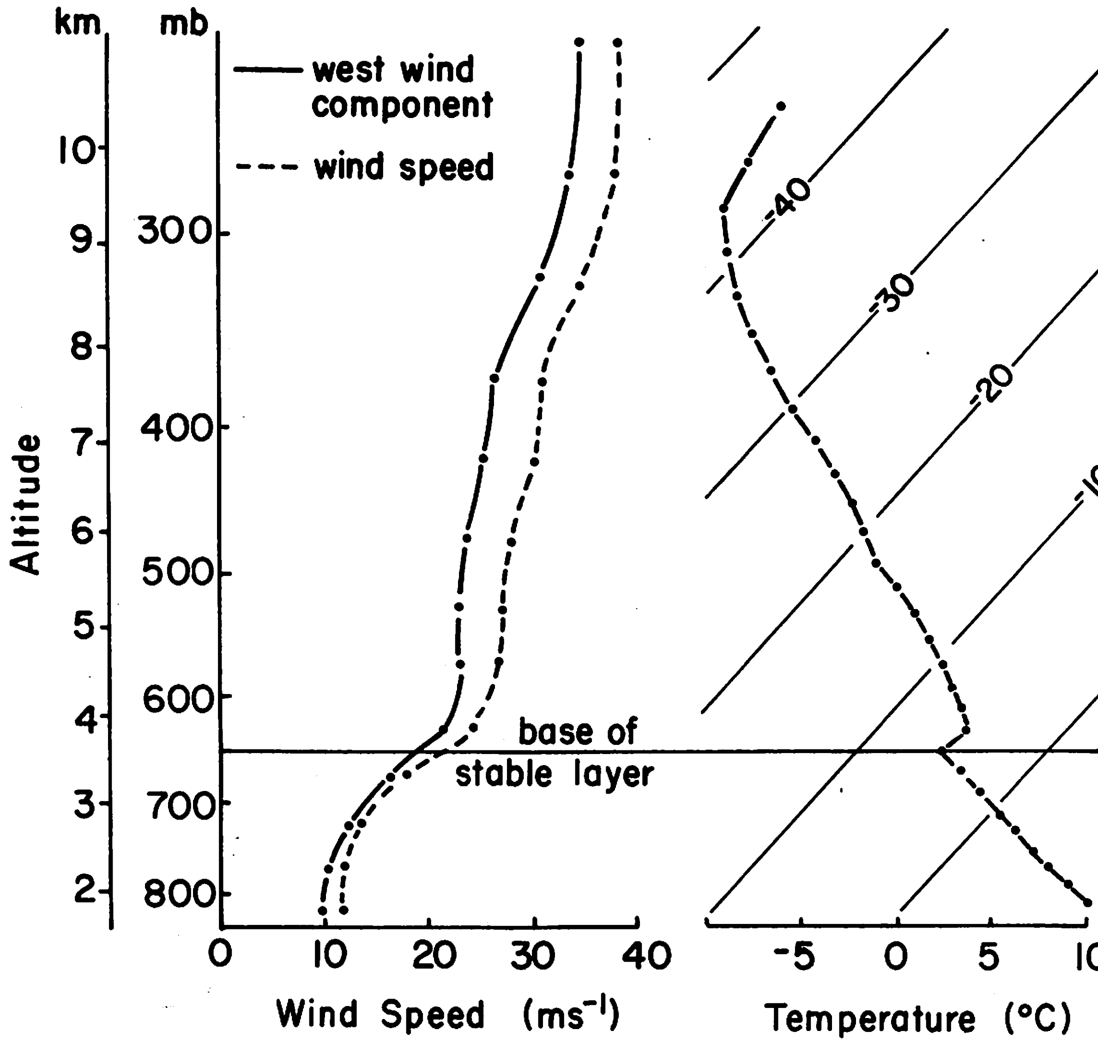


Fig. 13.7. Composite Denver sounding for time close to or during wind storms in Boulder.

Another difference between the upwind and downwind temperature profiles is the steeper lapse rate for the downwind sounding, with higher potential temperature at lower levels and lower temperatures above as compared with the upwind profile. The difference in lapse rate (i.e., 7.1°C km^{-1} vs. 5.7°C km^{-1}; statistically significant) is the result of horizontal convergence (divergence of streamlines in the vertical) with descending motion in the lower part of the disturbed flow and ascending motion above. Fig. 13.8 (top), taken from Lilly (1971), is an example of such flow pattern, showing the potential temperature cross section a few hours after one of the wind storms considered in the present study.

The cold air in the upwind sounding below 700 mb, which is not present over Denver, suggests that blocking of low level inversion layers windward of the mountains occurred in a number of cases. The reduced upflow on the windward slopes may, in turn, be one reason for the lack of precipitation on those slopes during storms in Boulder (see Section 11).

The location of the tropopause, just above 300 mb, is slightly higher than in the upwind sounding.

The major difference between the upstream and downstream wind profiles lies not in the mean shear across and below the stable layer (i.e., 7.5 m s^{-1} km^{-1} downwind vs. 6.4 m s^{-1} km^{-1} upwind; the difference is statistically not significant) but in the shear above it. In the middle and upper troposphere the wind speed continued to increase in the undisturbed flow upwind but remained remarkably constant through a deep layer downwind of the disturbance

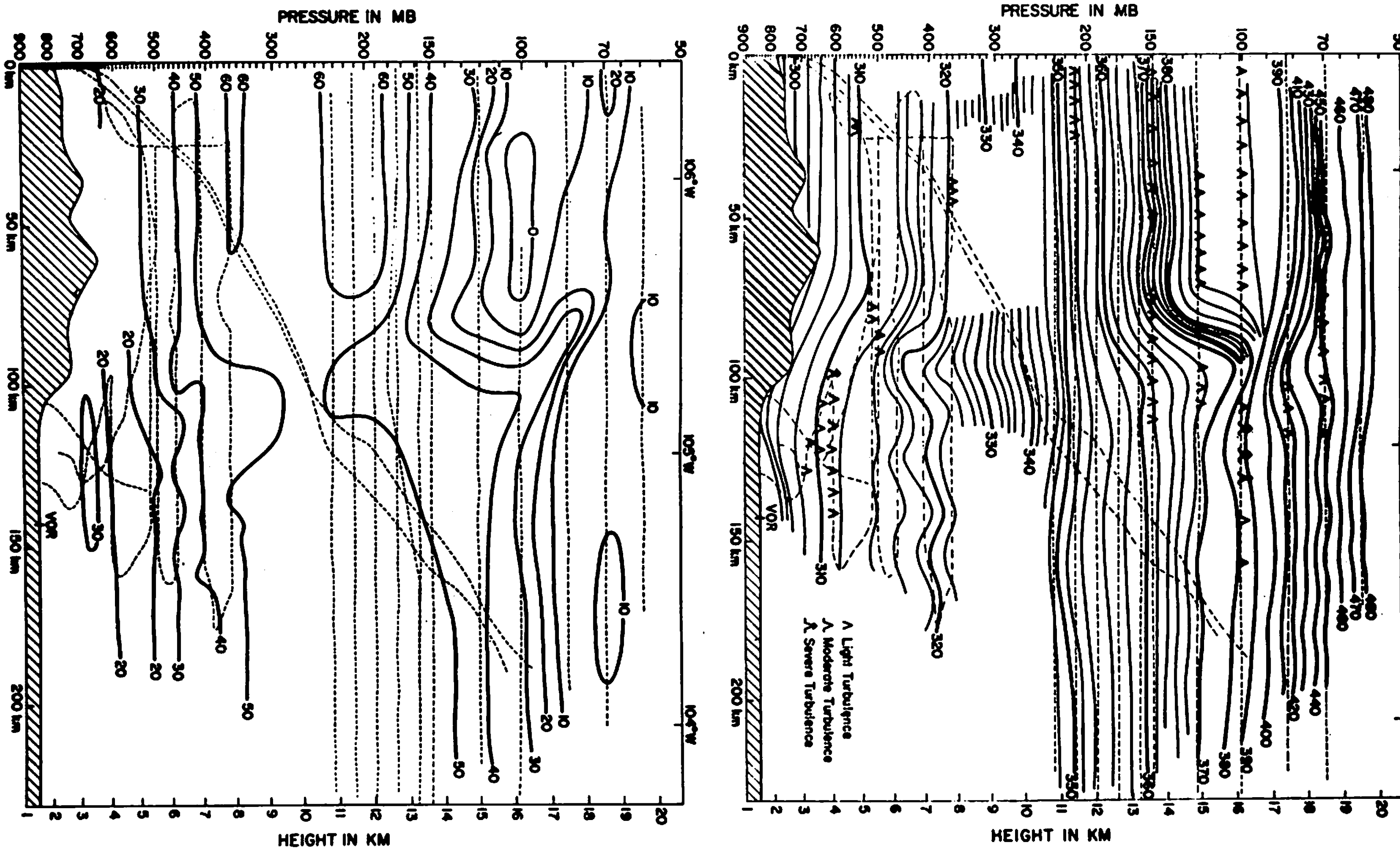


Fig. 13.8. Potential temperature cross section for February 17, 1970. Solid lines are isentropes, dashed lines are aircraft or balloon flight trajectories. The cross section is along a 275°-095° true azimuth line, crossing Kremmling, Colorado, and Denver VOR aircraft navigation stations (top); and westerly wind component cross section for February 17, 1970 (bottom), (from Lilly 1971).

(i.e., 1.2 m s^{-1} km^{-1} downwind vs. 3.6 m s^{-1} km^{-1} upwind; the difference is statistically significant). This is another result of streamline deformation in the disturbed flow. Fig. 13.8 (bottom), which is the cross section of west wind components corresponding to the potential temperature field in the same diagram, shows how this is reflected in a decrease in speed at upper levels (around 300 mb) and an increase below (around 700 mb).

For comparative purposes, mean wind speed profiles at Denver, and an indication of the variability, for the winters (December) 1947-1952 are shown in Fig. 13.9. Mean speed as well as west wind components at times of storms in Boulder were about one standard deviation (or a factor of two) above the 5-year mean below the stable layer, were three times the mean at that level, but were much less than that in the middle and upper troposphere (above 550 mb). With the downwind stable layer and associated strong winds being somewhat below mountain top levels, mean west wind components at mountain top are considerably greater downwind than upwind of the disturbance (i.e., 22.4 m s^{-1} vs. 18.6 m s^{-1}; the difference is significant) while there is little difference at the level of the stable layer (i.e., 22.4 m s^{-1} vs. 24.1 m s^{-1}).

Fig. 13.10 shows composite downwind soundings for the warm, cold and indifferent cases as defined in Section 8 (made up of 6, 13 and 13 soundings, respectively). The locations of the stable layer for the storm types were not found to be significantly different from the overall mean level, with the modal class located around 650 mb. <u>Comparison of the profiles suggests little</u>

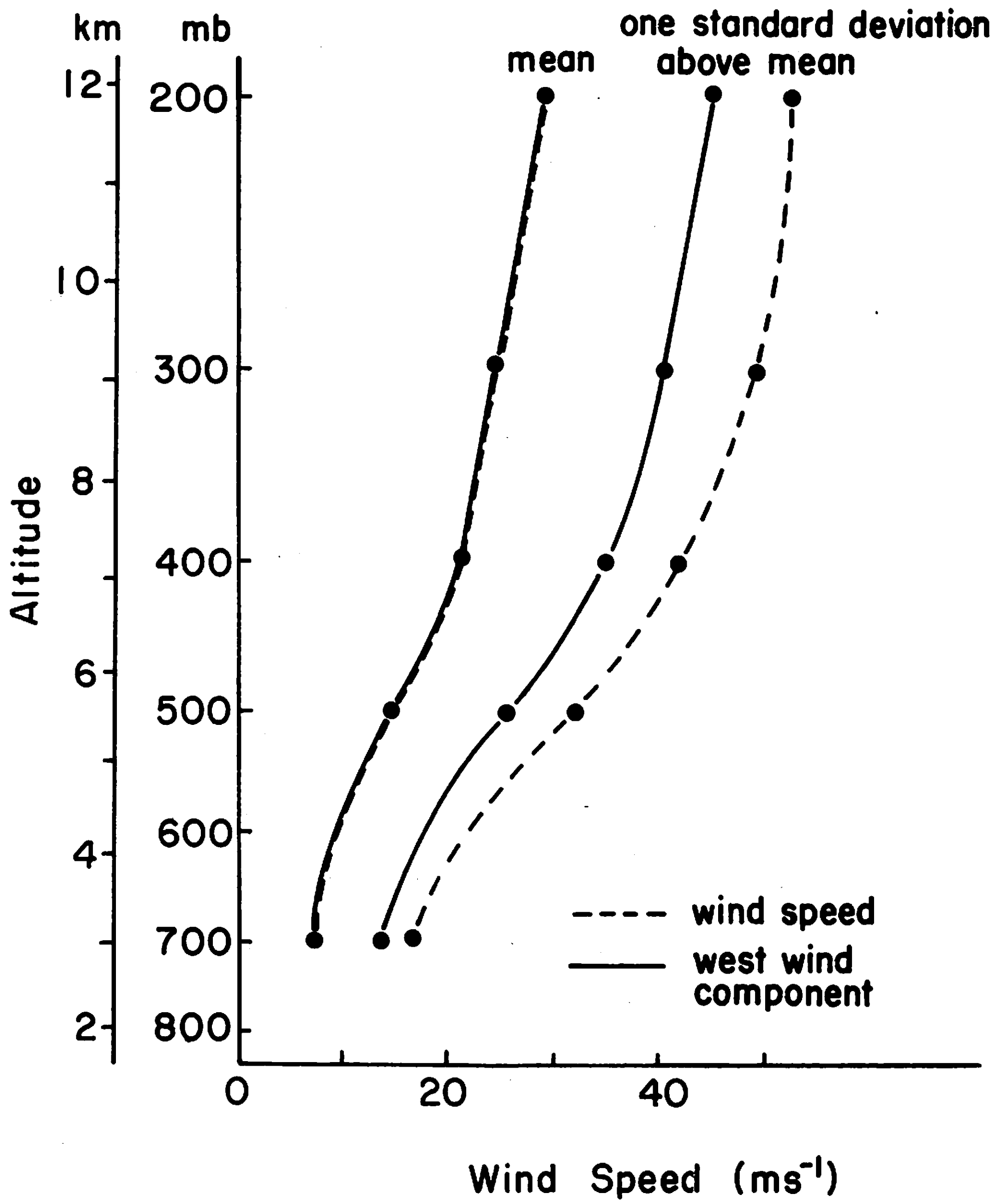


Fig. 13.9. Mean wind speed profile at Denver, winter (December) 1947-1952, data from Crutcher(1958).

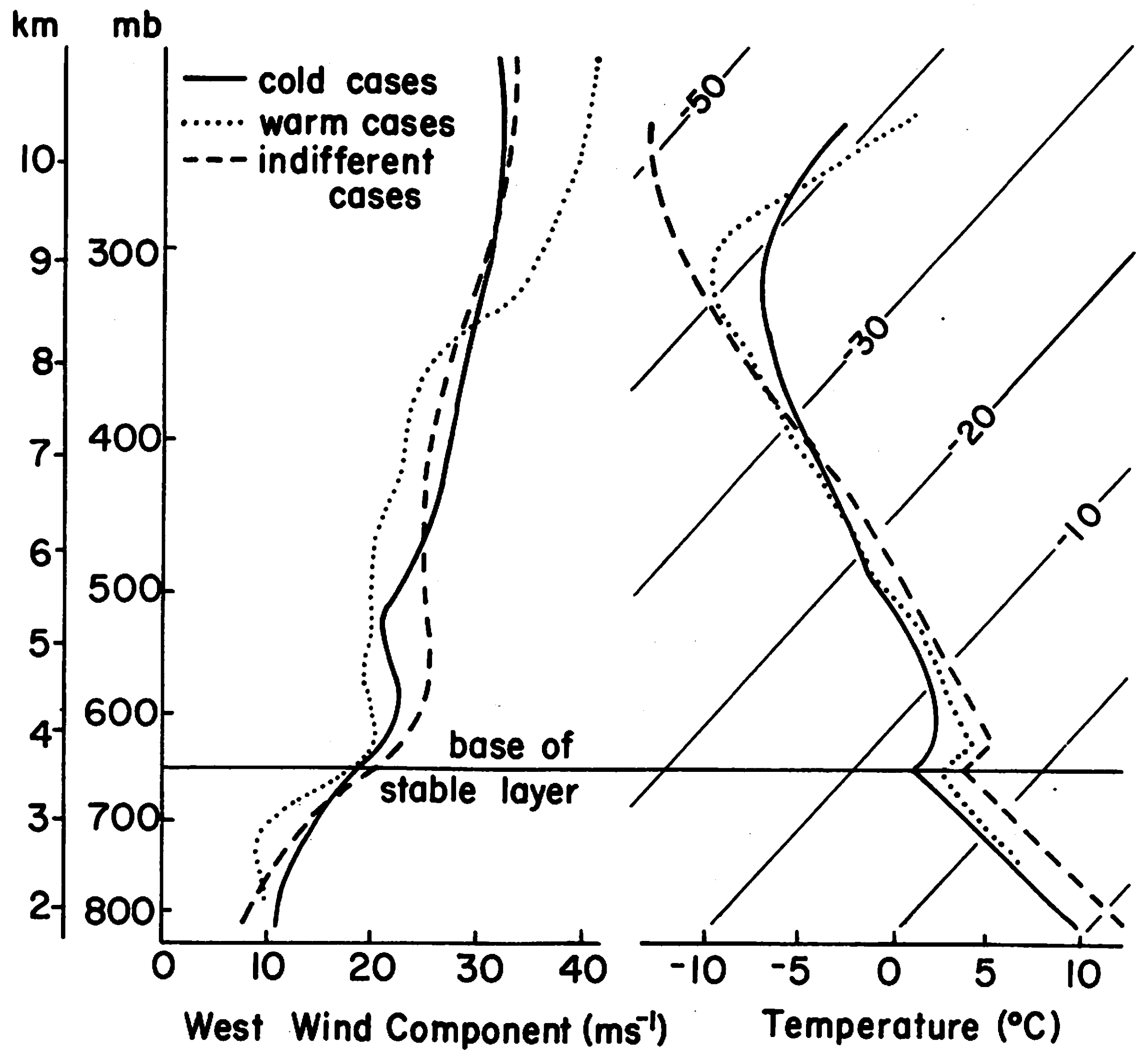


Fig. 13.10. Composite Denver sounding for time close to or during cold, warm and indifferent storms in Boulder.

<u>difference between the warm and cold cases</u>. The warm storms appear to have had the strongest wind shear below the stable layer and the indifferent ones the highest west wind components above it, the latter are also warmer than the storms defined as 'warm', but none of the differences are statistically significant. (The composite temperature profile for the warm cases could not be extended downward to the 800 mb level because of some very low stable layers, and the loss of even one temperature value out of 6 would have had a significant effect on the mean.)

Comparisons between the upwind and downwind profiles for the wind storm types show that the temperature of the air below the upwind and downwind stable layer is in close agreement (within about one degree) for the warm and indifferent cases. For the cold storms this is true only for the upper part of this air, the lower levels are much colder upstream which suggests that blocking of low level inversion layers windward of the mountains is mainly a feature of the cold storms.

The composite profile for the 32 soundings shown in Fig. 13.7 is not quite representative of conditions during wind storms since it still consists of soundings taken during as well as before or after the actual storm period in Boulder. To separate them, 'Boulder storm' soundings were considered to be those taken at least two hours after the onset or before the end of wind storms in Boulder while 'slope wind' soundings had to be taken at least two hours before the onset or after the end of storms and in addition westerly winds at Gold Hill and/or C-1 had to be ≥ 8 m s^{-1}.

This selection resulted in 6 'Boulder storm' soundings for which the base of the stable layer occurred most frequently around 675 mb. For the 7 'slope wind' soundings this level showed a very wide spread and an indication of a bimodal distribution; the median was found to be at 625 mb. The stable layer was therefore about 50 mb lower during 'Boulder storms' as compared with 'slope winds' (significant at the 10 per cent level). This may be interpreted to reflect a lower initial level in the undisturbed flow during storms in Boulder as indicated in Fig. 9.22, for instance.

The two composite profiles of temperature and west wind component are plotted in Fig. 13.11. Comparison of the results shows that with the more precise definition the 'Boulder storm' cases differ from 'winds on the slopes' in having a stronger west wind component just above the stable layer (25.1 m s^{-1} vs. 17.6 m s^{-1}; the difference is significant) resulting in a larger wind shear (10.4 m s^{-1} km^{-1} vs. 4.4 m s^{-1} km^{-1}; statistically significant) across and below the stable layer. The thickness of this layer is about the same (i.e., 20-25 mb) while the lapse rate above it is steeper for the 'Boulder storm' cases (i.e., 8.1°C km^{-1} vs. 7.5°C km^{-1}; the difference is significant). This is presumably the result of greater vertical divergence, which is also suggested by the low level wind maximum just above the stable layer. (Since the velocity decrease above the maximum occurs not only in the west wind component but also in the mean speed profile, it is not just the result of a shift in wind direction.) Moreover, this 'low level jet' cannot be used to explain the high surface wind in Boulder since mean gusts at the Southern Hills station during

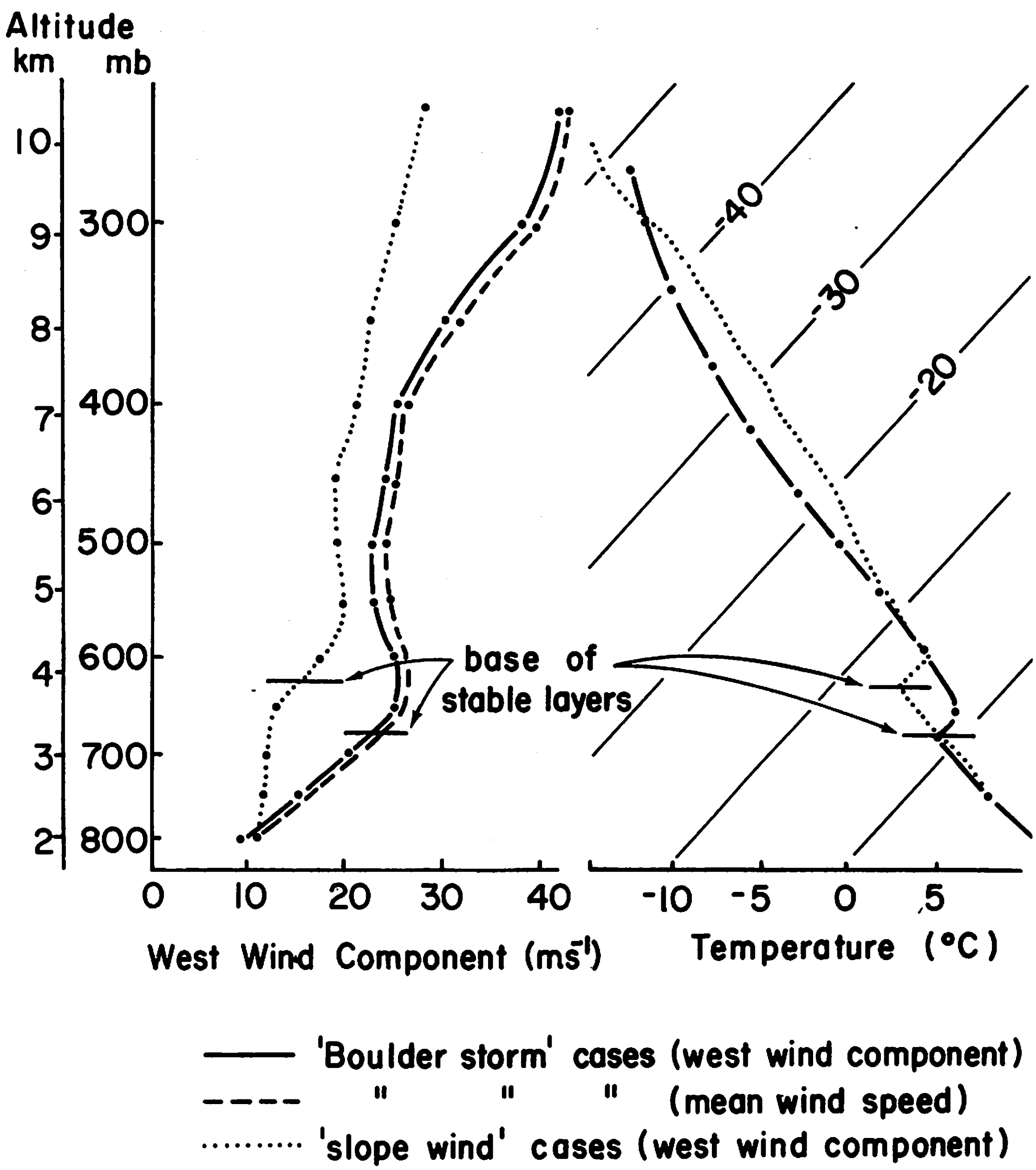


Fig. 13.11. Composite Denver sounding for time of storms in Boulder and wind on the slopes.

wind storms (i.e., 36 m s^{-1}) were much higher than the mean wind above Denver up to the 350 mb level.

Comparison of the 'Boulder storm' profile with the mean upwind sounding shows that the major differences above the stable layer are much steeper lapse rate and lower wind shear for the 'Boulder storm' cases (both are statistically significant). The extremely low level of the stable layer during 'Boulder storms' (the difference of 100 mb is statistically significant), but similar wind speeds at that level, caused the wind shear in the Denver sounding to be concentrated in a very shallow layer and thus increased it significantly to a value equivalent to that found near the jet stream (i.e., 10.1 m s^{-1} km^{-1} vs. 6.4 m s^{-1} km^{-1}; the difference is significant). This seems to indicate that <u>an extremely strong wind shear at lower levels is a characteristic of disturbed flow rather than an undisturbed one</u>.

Regarding wind direction, Fig. 13.12 shows its variation in the vertical for all 32 soundings and for the 6 'Boulder storm' cases. For all soundings the mean direction from 700 mb up to at least the 350 mb level was between 280 and 290°, with a variation (one standard deviation) between about 265° and 315°. For the 'Boulder storms', the wind direction above the level of the stable layer was not much different from the mean of the 32 soundings except for the smaller deviations, only below that level was there some backing to 270°. These results are rather interesting since the main mountain range in eastern Colorado trends north-south, yet, <u>the mean wind direction above mountain top levels at times of</u>

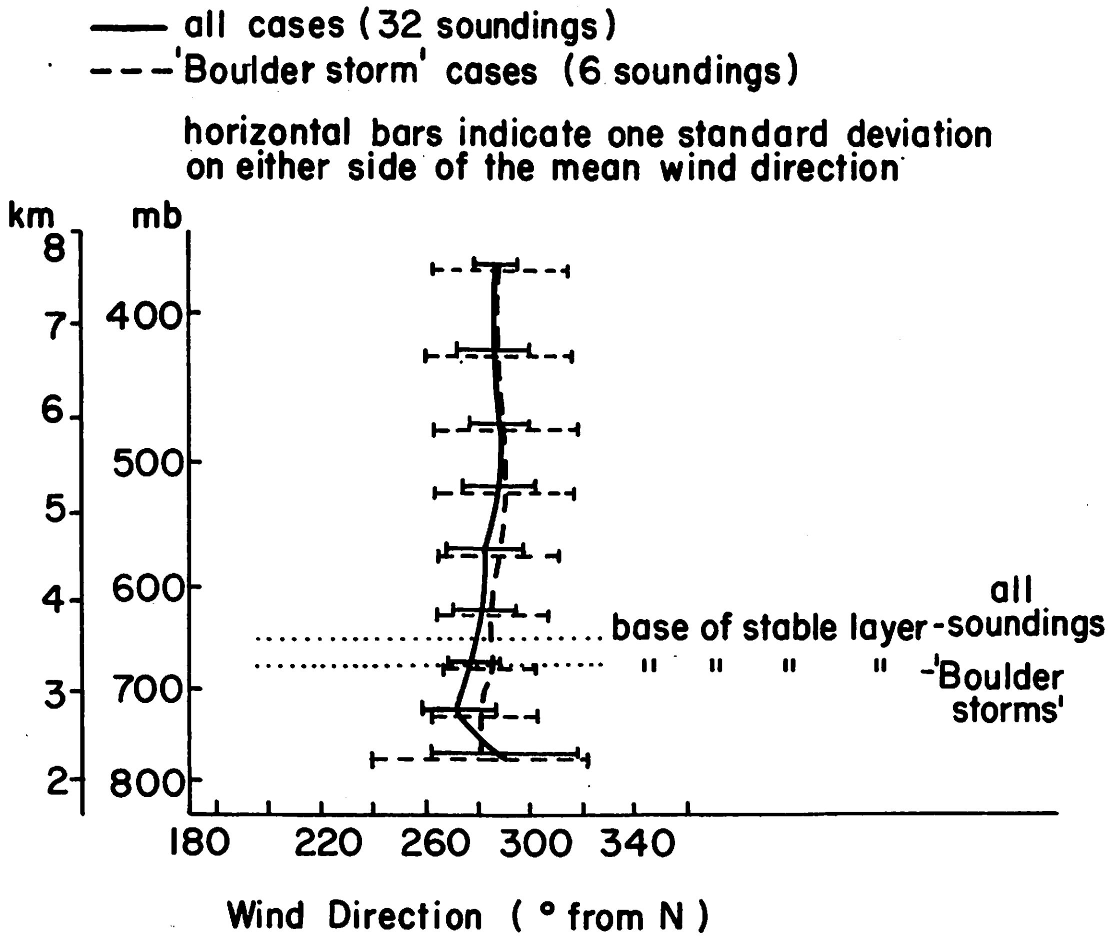


Fig. 13.12. Composite wind direction profile at Denver for all soundings (Fig. 13.7) and for 'Boulder storms' (Fig. 13.11).

<u>wind storms was not westerly, as one might expect, but somewhat north of west</u>. Mountain waves over eastern Colorado, too, have been found to be more likely with a northwesterly wind. MacDonald and Harrison (1960) suggested that the explanation may be the latitudinal variation in the Coriolis parameter: Colson (1950) has shown that a lee pressure trough is deepest when the air flow across a north-south mountain range is from the northwest as compared with other westerly directions; this would increase the surface pressure gradient across the mountain range (and thus make conditions for wind storms more favourable). This explanation may, however, be more applicable to geostrophic flow than meso-scale waves. An alternative explanation suggested by Lilly (1973, pers. comm.), may be the many mountain ranges to the southwest removing momentum from air flow coming from that direction.

c. Summary. During storm periods in Boulder the air flow across the mountains was found to be characterized by a well defined stable layer located slightly above mountain top levels (at about 575 mb) in the undisturbed flow. Downstream, at Denver, the altitude of this layer was found below mountain top levels (at about 650 mb). A difference in the downstream level of this stable layer was identified between 'slope winds' (at about 625 mb) and 'Boulder storms' (at about 675 mb), indicating a downslope/upslope movement of the surface wind maximum in response to a lowering/raising of this layer. Such critical importance of the altitude of this stable layer is in agreement with theoretical results (Section 3) and its recognition should thus significantly improve the possibility of forecasting Boulder's wind storms.

The mean depth of the stable layer was not found to be affected by the disturbance in the flow, but its altitude, as well as wind shear and lapse rate above and below it, were significantly altered between upstream and downstream soundings. Results derived from analysis of solely downwind data should therefore be interpreted with these limitations in mind.

Upstream and downstream winds in the upper troposphere were found to be generally less than 40 m s^{-1}; high winds at upper levels are therefore not necessary requirements for storms in Boulder. Furthermore, mean gust speeds at the surface in Boulder in excess of mid-tropospheric velocities in the upwind and downwind soundings show that strong surface winds at the foot of the mountains are not the result of an 'upper tropospheric jet' coming down to the ground.

The mean wind direction of the air flow up to the tropopause was found to be somewhat north of west, rather than west; but the reason for this is not yet fully understood.

During the 20 wind storm cases blocking of low level inversion layers on the windward side was found to be important, mainly with the cold storms, which would partly explain the insignificant amounts of precipitation west of the Divide found in Section 11 although subsidence in the northwesterly air stream is probably of equal importance.

Cold, warm and indifferent wind stormsdiffered somewhat in the windward sounding but were quite similar downwind of the disturbance. In many respects the 'indifferent' cases appeared to be more closely related to the 'warm' storms than the 'cold' ones.

CHAPTER IV

A SYNTHESIS

14. Boulder's Winds - Boras and Chinooks

Boulder's wind storms are not all Chinooks as they are frequently referred to in the literature, but in fact cover the whole range of downslope wind types, from bora to cold air foehn to foehn. Analysis showed that of the 21 cases (20 storms and one additional phase), 4 cases (19%) were genetically warm, 8 cases (38%) were classified as indifferent but their characteristics suggest that many of these may have been warm, and 9 cases (43%) were found to have been genetically cold. However, an increasing number of the genetically cold storms appeared to be warm with decreasing elevation because of removal of surface inversions or supression of their development and diabatic warming. Thus, only 4 of the genetically cold storms appeared to be cold in Boulder and only two in Denver. However, the few cold type storms that arrived still cold at Boulder and Denver produced temperature decreases equivalent to those observed with the classic Bora of Yugoslavia. This is remarkable considering the fact that part of the cold effect of the Bora is attributable to the contrast provided by the warm waters of the Adriatic. Boulder's warm storms, on the other hand, produced temperature increases comparable with those observed in other Chinook wind locations and greater than many in the classic Foehn area north of the European Alps. However, the warmth of Boulder's warm storms is not the result of the classic 'foehn effect' because precipitation on the windward slopes was found to be insufficient for it in any of the storm cases and must therefore be the result of warm air advection and subsidence.

The most important difference between Boulder's warm and cold storms was found to be the surface pressure trend because of differences in the synoptic situation leading to the two types of storms. In all other respects both types were very similar.

Whether the occurrence of genetically cold winds as well as warm ones in the same locality is unique is difficult to determine from the more or less descriptive discussions of the temperature characteristics of other downslope winds. The apparently outstanding characteristics of Boulder's storms are not unique, however. Their violence and extreme gustiness, the very narrow belt affected by the high winds, the interruptions or pauses, the fluctuations in surface pressure, are features common to cold and warm downslope winds in other parts of the world. Since in the case of the Boulder storms it was shown that most of these characteristics are related to the development and movement of a 'lee flow disturbance', this would suggest that perhaps all downslope wind storms are generated by a similar mechanism (this does not include, of course, marked changes in temperature without accompanying high winds which are not considered here).

15. Suggestion for the Development of a Forecast Aid for Boulder's Storms

Attempts have been made in the past to 'objectively' forecast Boulder's wind storms using statistical techniques (Sangster 1970; 1972). The most recent work (Sangster 1972) suggests that the highest probability for strong winds exists with high west wind components at the 500 mb level at Salt Lake, Grand Junction and

Lander, and a large downward slope in the 850 and 700 mb constant pressure surfaces between Salt Lake/Grand Junction and Lander. The results, particularly with respect to the prediction of very high winds, are not very encouraging, however.

Based on the results of the present study, it is suggested that important predictors are the temperature stratification above mountain top levels (showing a stable layer) as well as wind speeds at that level, i.e., around 600 mb (quite strong: 20-30 m s^{-1} at the upwind station), and in the upper troposphere, i.e., around 300 mb (not too high: only 30-40 m s^{-1} at the upwind station). -- The fact that high wind speeds in the upper troposphere are not a necessary requirement for strong downslope winds is certainly shown by the occurrence of the Santa Ana of Southern California, the Wasatch wind of Utah and the Bora of Yugoslavia on the western slopes of mid-latitude mountain ranges where winds at upper levels are rarely easterly. -- Surface pressure trends are not very helpful since both, decreases and increases, may lead to storms; but a rapidly increasing pressure gradient across the mountains could be used as a warning sign.

16. The Night the Wind Blew

In the following section the major findings of the present study are synthesized. This is done in form of a 'scenario' which describes the physical events prior to and during an average wind storm in Boulder which have been found to be of importance. It is not a complete description of what happens during a storm, certain processes are still incompletely understood and need further investigation (e.g., Where do the stable layers associated with the warm and indifferent storms come from? Is subsidence perhaps an important mechanism in these cases?) Nevertheless, this study has resulted in enough pieces of information to describe the most important sequence of events.

"In the afternoon of a day in January surface pressure trends at stations across Colorado suddenly change signs.

COLD STORMS	WARM/INDIFFERENT STORMS
A cold front crossing Colorado from west to east causes surface pressure to rise west of the Divide and a few hours later the change in trend is also observed in the lee. The pressure rises, however, twice as fast on the windward side as in the lee, creating an increasing pressure gradient across the mountains. The cold air behind the cold front causes temperatures along the lee slopes to decrease but because of a strong surface inversion the effect is a warming at the foot of the mountains.	A low pressure system to the north/ northwest with a front extending southeastward across Colorado causes surface pressures to decrease at stations on either side of the Divide. The drop is, however, twice as rapid in the lee as on the windward side, creating an increasing pressure gradient across the mountains. There may or may not be a rise in temperature at the lee stations on the slopes but because of a surface inversion a warming is observed at the foot of the mountains.

Several hours later at the upwind station west of the Divide a stable layer with its base at about 575 mb (about 750 m above mountain top levels) is observed. Just below it, a west wind component of about 20 m s^{-1} is recorded. Wind speeds throughout most of the layer below the stable region are much higher than normal while those in the upper troposphere are not quite that unusual. East of the Divide, above Denver, the base of the stable layer is located at 625 mb (slightly above the mountains as a result of a lee disturbance developing in the air flow). Below, a west wind component of about 12 m s^{-1} is observed.

Over the upper lee slopes a deep 'lee flow disturbance' is now situated. Below it a local pressure minimum has developed toward which air is accelerated. Thus, winds at the mountain stations are already blowing, mean speeds reach 12-14 m s^{-1} with gusts up to 24-28 m s^{-1}.

Shortly before midnight some upstream air flow parameters (i.e., level of the stable layer/wind speed profile) are gradually shifting across critical threshold values, the 'lee flow disturbance' begins to move down the slopes and winds start to decrease at the mountain stations. At the foot of the mountains, a rapid drop in surface pressure, superimposed upon the general pressure trend, is observed; and then the winds begin to blow in Boulder. The Denver sounding shows some changes now in the structure of the air flow: the base of the stable layer is lower, at 625 mb, and the west wind component below it has doubled, now reading 24 m s^{-1}. Air is crossing the Divide from a direction slightly north of west (280-290°), subsidence within this air/some blocking on the windward side result in little, if any, precipitation west of the Divide, but a crest cloud covers the mountain tops.

After midnight, at the height of the storm in Boulder, the pressure gradient between Eagle and Denver is 8 mb, and a local pressure minimum of 5 mb has developed. Air accelerating down the slopes toward the pressure minimum increases in speed from 8-10 m s^{-1} and gusts from 20 m s^{-1} along the slopes to mean speeds of 20 m s^{-1} and to gusts of 36 m s^{-1} in Boulder. The high and extremely gusty winds at the foot of the mountains are, however, very localized; they decrease again rapidly by 50 per cent over a distance of about 10 km to the east into the plains, in agreement with the adverse pressure gradient between Boulder and Denver. At Denver, where mean speeds barely reach 5 m s^{-1}, people find it hard to believe that roofs are taking off in Boulder.

During the night, slight changes in upstream air mass characteristics cause small changes in the position of the 'lee air flow disturbance' which results in large variations in wind speed at a station near the edge of the phenomenon; at the other stations no major fluctuations are observed.

Later, winds in Denver also begin to pick up, reaching 10 m s^{-1}, strong enough to remove the shallow surface inversion layer and causing temperature rises larger than those observed earlier in Boulder. At that time, winds in Boulder are already beginning to decrease. Speeds increase at the mountain stations as the 'lee flow disturbance' moves upslope. Fortunately, upstream air flow characteristics seem to have deteriorated beyond the point optimum for wind storms in Boulder. --Another windy night may be over."

However, sometimes (given the right changes in the critical air flow characteristics, probably through some diurnal control such as diurnal changes in upwind lapse rate):

"the 'lee flow disturbance' moving upslope produces only brief pauses in Boulder and a short increase in surface winds at the lower mountain stations. A little after sunrise it moves back down again, bringing the second phase of the wind storm which 'creates some more havoc' in Boulder before noon."

BIBLIOGRAPHY

BIBLIOGRAPHY

Aanensen, C. J. M., 1964, Gales in Yorkshire in February 1962, Geophysical Memoirs, 14(108):1-44

Alaka, M. A. (ed.), 1960, The airflow over mountains, Technical Note No. 34, World Meteorological Organization, Geneva.

Arakawa, S., 1969, Climatological and dynamical studies on the local strong winds, mainly in Hokkaido, Japan, Geophysical Magazine, 34(4):359-425.

Ashwell, I., 1971, Warm blast across the snow-covered prairie, Geographical Magazine, 43(12):858-863.

Ashwell, I. Y., and Marsh, J. S., 1967, Moisture loss under chinook conditions, Proceedings: First Canadian Conference on Micrometeorology, Part 2, pp. 307-314, Meteorological Branch, Canada Department of Transport, Toronto.

Ball, F. K., 1956, The theory of strong katabatic winds, Australian Journal of Physics, 9(3):373-386.

Band, G., 1951, Die Bora der Adria, Geofisica Pura e Applicata, 19(3-4):186-219.

Barry, R. G., 1972, Climatic environment of the east slope of the Colorado Front Range, Occasional Paper No. 3, Institute of Arctic and Alpine Research, University of Colorado, Boulder.

Barry, R. G., and Chorley, R. J., 1968, Atmosphere, Weather and Climate, Methuen & Co. Ltd., London.

Beran, D. W., 1967, Large amplitude lee waves and chinook winds, Journal of Applied Meteorology, 6(2):865-877.

Berg, H., 1952, Der Foehn im noerdlichen Alpen- und Voralpengebiet vom 11. bis 14. Maerz 1951, Bericht des Deutschen Wetterdienst, 35:117-124.

Berry, F. A., Bollay, E., and Beers, N. R., 1945, Handbook of Meteorology, McGraw-Hill Book Company Inc., London.

Billwiller, R., 1899, Ueber verschiedene Entstehungsarten und Erscheinungsformen des Foehns, Meteorologische Zeitschrift, 16:204-215.

Binnie, A. M., and Orkney, J. C., 1955, Experiments on the flow of water from a reservoir through an open horizontal channel, II. The formation of hydraulic jumps, Proceedings of the Royal Society of London, A, 230(1181):237-246.

Blackadar, A. K., 1960, A survey of wind characteristics below 1500 feet, Meteorological Monographs, 4(22):3-11.

Bluethgen, J., 1966, Allgemeine Klimageograpie, W. de Gruyter, Berlin.

Borisov, A. A., 1965, The climates of the USSR, Aldine Publishing Co., Chicago.

Brinkmann, W. A. R., 1969a, A geography of foehn-type winds, Unpublished paper, Institute of Arctic and Alpine Research, University of Colorado, Boulder.

Brinkmann, W. A. R., 1969b, The definition of the chinook in the Calgary area, Unpublished M.Sc. thesis, Department of Geography, University of Calgary, Calgary.

Brinkmann, W. A. R., 1971, What is a foehn?, Weather, 26(6):230-239.

Brinkmann, W. A. R., and Zipser, E. J., 1972, Observational studies of strong downslope windstorms, Abstract of paper presented at the Fourth Conference on Weather Forecasting and Analysis, American Meteorological Society, Bulletin, 53(1):88.

Cadez, M., 1953, Ueber einige Einfluesse orographischer Hindernisse auf die Luftbewegung, Archiv fuer Meteorologie, Geophysik und Bioklimatologie, A, 6(3-4):403-416.

City of Boulder, 1970, Analysis of damage to residential properties from high winds occurring on January 7 and January 31, 1969, in Boulder, Colorado, City Manager's Office, Boulder.

Colson, D. V., 1950, Effects of a mountain range on quasi-stationary waves, Journal of Meteorology, 7(4):279-282.

Conrad, V., 1936, Die klimatologischen Elemente und ihre Abhaengigkeit von terrestrischen Einfluessen, in: Borchardt, W., et al., Handbuch der Klimatologie, Verlag von Gebrueder Borntraeger, Berlin.

Cook, A. W., and Topil, A. G., 1952, Some examples of chinooks east of the mountains in Colorado, American Meteorological Society, Bulletin, 33(2):42-47.

Corby, G. A., 1954, The airflow over mountains (A review of the state of current knowledge), Royal Meteorological Society, Quarterly Journal, 80(346):491-521.

Crutcher, H. L., 1958, Upper wind distribution statistical parameter estimates, Weather Bureau Technical Paper No. 34, United States Department of Commerce, Washington.

Danielsen, E. F., and Bleck, R., 1970, Tropospheric and stratospheric ducting of stationary mountain lee waves, Journal of the Atmospheric Sciences, 27(5):758-772.

Davenport, A. G., 1961, The spectrum of horizontal gustiness near the ground in high winds, Royal Meteorological Society, Quarterly Journal, 87(372):194-211.

Davis, F. K., and Newstein, H., 1968, The variation of gust factors with mean wind speed and with height, Journal of Applied Meteorology, 7(3):372-378.

Deacon, E. L., 1955, Gust variation with height up to 150 m, Royal Meteorological Society, Quarterly Journal, 81(350):562-573.

Defant, F., 1951, Local winds, in: Malone T. F. (ed.), Compendium of Meteorology, pp. 655-672, American Meteorological Society, Boston.

Eckert, G., and Klinkow, W., 1947, Der "Vardar"-Wind der Nordwest-Aegaeis, Zeitschrift fuer Meteorologie, 1(11-12):365-369.

Ficker, H. von, and Rudder, B. de, 1948, Foehn und Foehnwirkungen: Der gegenwaertige Stand der Frage, Probleme der Bioklimatologie, 1:1-114.

Fiedler, F., and Panofsky, H. A., 1970, Atmospheric scales and spectral gaps, American Meteorological Society, Bulletin, 51(12):1114-1119.

Fleagle, R. G., 1950, A theory of air drainage, Journal of Meteorology, 7(3):227-232.

Foerchtgott, J., 1957, Active turbulent layer downwind of mountain ridges, Schweizer Aero-Review, 32:324-335.

Freeman, J. C., 1948, An analogy between the equatorial easterlies and supersonic gas flow, Journal of Meteorology, 5(4):138-146.

Freeman, J. C., 1951, The solution of nonlinear meteorological problems by the method of characteristics, in: Malone, T. F. (ed.), Compendium of Meteorology, pp. 421-433, American Meteorological Society, Boston.

Frey, K., 1953, Die Entwicklung des Sued- und des Nordfoehns, Archiv fuer Meteorlogie, Geophysik und Bioklimatologie, A, 5(4):432-477.

Frey, K., 1957, Zur Diagnose des Foehns, Meteorologische Rundschau, 10(6):181-185.

Fujita, T., 1955, Results of detailed synoptic studies of squall lines, Tellus, 7(4):405-436.

Gentilli, J., and Fairbridge, R. W., 1967, Winds - Local, in: Fairbridge, R. W. (ed.), The Encyclopedia of Atmospheric Sciences and Astrogeology, pp. 1151-1156, Reinhold Publishing Corporation, New York.

Georgii, W., 1952, Beitrag zur Aerologie von Mendoza, Deutscher Wetterdienst in der US-Zone, 36:1-12.

Georgii, W., 1954, Studie ueber den Zondawind der Kordillere von Mendoza, Meteorologische Rundschau, 7(7-8):125-129.

Georgii, W., 1967, Thermodynamik und Kinematik des Kaltluftfoehns, Archiv fuer Meteorologie, Geophysik und Bioklimatologie, A, 16(2-3):137-152.

Gerbier, N., and Berenger, M., 1961, Experimental studies of lee waves in the French Alps, Royal Meteorological Society, Quarterly Journal, 87(371):13-23.

Gill, G. C., 1967, On the dynamic response of meteorological sensors and recorders, Proceedings: First Canadian Conference on Micrometeorology, Part 1, pp. 1-27, Meteorological Branch, Canada Department of Transport, Toronto.

Glenn, C. L., 1961, The chinook, Weatherwise, 14(5):175-182.

Grober, K. W., 1961, Die Bora und ihr landschaftsgestaltender Charakter, Geographische Rundschau, 13:321-323.

Hader, F., 1946, Der Foehn im Raum von Lunz am See am 27. Februar 1947, Geographische Gesellschaft Wien, Mitteilungen, 89(7-12): 142-143.

Hann, J., 1885, Einige Bemerkungen zur Entwickelungs-Geschichte der Ansichten ueber den Ursprung des Foehns, Meteorologische Zeitschrift, 2(11):393-399.

Hann, J. 1903, Handbook of Climatology, Part 1: General Climatology, MacMillan Co., New York.

Harrison, H. T., 1956, Synoptic features of the mountain wave at Denver, Colorado, Meteorology Circular No. 41, United Air Lines Inc., Denver.

Harrison, H. T., 1957, Forecasting the mountain wave at Denver, Colorado , Meteorology Circular No. 42, United Air Lines Inc., Denver.

Henson, W. R., 1952, Chinook winds and red belt injury to lodgepole pine in the Rocky Mountain parks area of Canada, Contribution No. 28, pp. 62-64, Division of Forest Biology, Canada Department of Agriculture, Ottawa.

Hoinkes, H., 1950, Foehnentwicklung durch Hoehentiefdruckgebiete, Archiv fuer Meteorologie, Geophysik und Bioklimatologie, A, 2(1):82-96.

Hoinkes, H., 1953, Probleme der alpinen Synoptic, Archiv fuer Meteorologie, Geophysik und Bioklimatologie, A, 5(3):304-325.

Holmboe, J., and Klieforth, H., 1957, Investigations of mountain lee waves and the air flow over the Sierra Nevada, Final Report No. AF 19(604)-728, Department of Meteorology, University of California, Los Angeles.

Houghton, D. D., and Isaacson, E., 1968, Mountain winds, Studies in Numerical Analysis, 2:21-52.

Houghton, D. D., and Kasahara, A., 1968, Nonlinear shallow fluid flow over an isolated ridge, Communications on Pure and Applied Mathematics, 21:1-23.

Ives, R. L., 1950, Frequency and physical effects of chinook winds in the Colorado High Plains region, Association of American Geographers, Annals, 40(4):293-327.

Jenkins, G. M., and Watts, D. G., 1968, Spectral Analysis, Holden-Day, San Francisco.

Johnson, H. N., 1936, The dry chinook wind, American Meteological Society, Bulletin, 17(1):23-27.

Julian, L. T., and Julian, P. R., 1969, Boulder's winds, Weatherwise, 22(3):108-112 & 126.

Kenney, J. F., and Keeping, E. S., 1954, Mathematics of Statistics, D. van Nostrand Company Inc., Princeton.

Klemp, J. B., and Lilly, D. K., 1973, A mechanism for the generation of downslope windstorms, Royal Meteorological Society, Quarterly Journal, to be submitted.

Kuettner, J., 1939, Zur Entstehung der Foehnwelle, Beitraege zur Physik der Freien Atmosphaere, 25:251-299.

Kuettner, J., 1940, Der Kosava in Serbien, Meteorologische Zeitschrift, 57(3):120-122.

Kuettner, J., 1959, The rotor flow in the lee of mountains, GRD Research Notes No. 6, Geophysics Research Directorate, Air Force Cambridge Research Center, Bedford.

Kuettner, J., and Lilly, D. K., 1968, Lee waves in the Colorado Rockies, Weatherwise, 21(5):180-185.

Lamb, P. J., 1970, An investigation of the Canterbury Nor'-wester, Unpublished M.A. thesis, Department of Geography, University of Canterbury, Christchurch.

Larsson, L., 1954, Observations of lee wave clouds in the Jaemtland Mountains, Sweden, Tellus, 6(2):124-138.

Lettau, H. H., and Haugen, D. A., 1960, Wind, in: Handbook of Geophysics, pp. 5/1-5/16, MacMillan Co., New York.

Lilly, D. K., 1971, Observations of mountain-induced turbulence, Journal of Geophysical Research, 76(27):6585-6588.

Lilly, D. K., 1972, Wave momentum flux - a GARP problem, American Meteorological Society, Bulletin, 53(1):17-23.

Lilly, D. K., Pann, Y., Kennedy, P., and Toutenhoofd, W., 1971, Data catalog for the 1970 Colorado lee wave observational program, Technical Notes NCAR-TN/STR-72, National Center for Atmospheric Research, Boulder.

Lilly, D. K., and Toutenhoofd, W., 1969, The Colorado lee wave program, in: Pao, Y., and Goldburg, A. (eds.), Proceedings: Clear Air Turbulence and its Detection, pp. 232-245, Plenum Press, New York.

Lilly, D. K., and Zipser, E. J., 1972, The Front Range windstorm of 11 January 1972, Weatherwise, 25(2):56-63.

Lockwood, J. G., 1962, Occurrence of foehn winds in the British Isles, Meteorological Magazine, 91(1076):57-65.

Loewe, F., 1950, A note on katabatic winds at the coasts of Adelieland and King George V Land, Geofisica Pura e Applicata, 16(3-4):159-162.

Long, R. R., 1954, Some aspects of the flow of stratified fluids, II. Experiments with a two-fluid system, Tellus, 6(2):97-115.

Longley, R. W., 1967, The frequency of winter chinooks in Alberta, Atmosphere, 5(4):4-16.

Lovill, J. E., 1969, Transport processes in orographically induced gravity waves as indicated by atmospheric ozone, Atmospheric Science Paper No. 135, Department of Atmospheric Science, Colorado State University, Fort Collins.

MacDonald, N. J., and Harrison, H. T., 1960, Some observations of the mountain wave in eastern Colorado, American Meteorological Society, Bulletin, 41(11):627-632.

MacHattie, L. B., 1963, Winter injury of lodgepole pine foliage, Weather, 18(10):301-307.

Madden, R. A., and Julian, P. R., 1971, Detection of a 40-50 day oscillation in the zonal wind in the tropical Pacific, Journal of the Atmospheric Sciences, 28(5):702-708.

Manley, G., 1945, The helm wind of Crossfell, 1937-1939, Royal Meteorological Society, Quarterly Journal, 71(309-310):197-219.

Marsh, J. S., 1965, The chinook and its geographic significance in southern Alberta, Unpublished M.Sc. thesis, Department of Geography, University of Alberta, Calgary.

Middleton, W. E. K., and Spilhaus, A. F., 1953, Meteorological Instruments, University of Toronto Press, Toronto.

Miller, D. J., 1972, Human perception of and adjustment to the high wind hazard in Boulder, Colorado, Unpublished M.A. thesis, Department of Geography, University of Colorado, Boulder.

Mirrlees, S. T. A., 1934, Meteorological results of the British Arctic air-route expedition 1930-1931, Geophysical Memoirs, 61:1-61.

Mitchell, J. M., 1966, Climatic change, Technical Note No. 79, World Meteorological Organization, Geneva.

Newton, C. W., 1971, Mountain torques in the global angular momentum balance, Journal of the Atmospheric Sciences, 28(4): 623-628.

Obenland, E., 1956, Untersuchung zur Foehnstatistik des Oberallgaeus, Bericht des Deutschen Wetterdienst, 23:1-40.

Paddock, M. W., 1964, The climate and topography of the Boulder region, in: Rodeck, H. G. (ed.), Natural History of the Boulder Area, pp. 25-33, Leaflet No. 13, University of Colorado Museum, University of Colorado, Boulder.

Panofsky, H. A., and Van Der Hoven, I., 1955, Spectra and cross-spectra of velocity components in the mesometeorological range, Royal Meteorological Society, Quarterly Journal, 81(350): 603-606.

Quenouille, M. H., 1952, Associated Measurements, Butterworths Scientific Publications, London.

Ratner, B., 1959, Upper-air climatology of the United States, Weather Bureau Technical Paper No. 32, United States Department of Commerce, Washington.

Reiter, E., 1969, Local circulation systems: Foehn and katabatic winds, in: Rex, D. F. (ed.), World Survey of Climatology, Volume 4, pp. 188-190, Elsevier Publishing Company, Amsterdam.

Riehl, H., 1971, An unusual chinook case, Weather, 26(6):241-246.

Rossmann, F., 1950, Ueber das Absteigen des Foehns in die Taeler, Bericht des Deutschen Wetterdienst in der US-Zone, 12:94-98.

Rumney, G. R., 1968, Climatology and the World's Climates, The MacMillan Company, New York.

Ruskin, R. E., 1963, The measurement of humidity in meteorology, Weatherwise, 16(2):55-61 & 76.

Sangster, W. E., 1970, A forecast aid for Boulder winds, Technical Memorandum WBTM CR-31, Weather Bureau, United States Environmental Science Services Administration, Kansas City.

Sangster, W. E., 1972, An objective forecast technique for Colorado downslope winds, Technical Memorandum NWS CR-50, National Weather Service, United States National Oceanic and Atmospheric Administration, Kansas City.

Schmitt, W., 1930, Foehnerscheinungen und Foehngebiete, Wissenschaftliche Veroeffentlichungen des Deutschen und Oesterreichischen Alpenvereins No. 8, Verlag des Deutschen und Oesterreichischen Alpenvereins, Innsbruck.

Schweitzer, H., 1953, Versuch einer Erklaerung des Foehns als Luftstroemung mit ueberkritischer Geschwindigkeit, Archiv fuer Meteorologie, Geophysik und Bioklimatologie, A, 5(3):350-371.

Scorer, R. S., 1949, Theory of waves in the lee of mountains, Royal Meteorological Society, Quarterly Journal, 75(323):41-56.

Scorer, R. S., 1953, Theory of air-flow over mountains, II. The flow over a ridge, Royal Meteorological Society, Quarterly Journal, 79:70-83.

Scorer, R. S., and Klieforth, H., 1959, Theory of mountain waves of large amplitude, Royal Meteorological Society, Quarterly Journal, 85(364):131-143.

Sergius, L. A., 1952, The Santa Ana, Weatherwise, 5(3):66-68.

Sergius, L. A., Ellis, G. R., and Ogden, R. M., 1962, The Santa Ana winds of southern California, Weatherwise, 15(3):102-105 & 121.

Snedecor, G. W., and Cochran, W. G., 1967, Statistical Methods, The Iowa State University Press, Ames.

Sutton, O. G., 1953, Micrometeorology, McGraw-Hill Book Company Inc., New York.

Tepper, M., 1950, A proposed mechanism of squall lines: The pressure jump line, Journal of Meteorology, 7(1):21-29.

Turner, R. W., 1966, Pincher Creek, Weather, 21(11):412-413.

Tyson, P. D., 1968, Nocturnal local winds in a Drakensberg Valley, The South African Geographical Journal, 50:15-32.

Van Der Hoven, I., 1957, Power spectrum of horizontal wind speed in the frequency range from 0.0007 to 900 cycles per hour, Journal of Meteorology, 14(2):160-164.

Vergeiner, I., 1971, An operational linear lee wave model for arbitrary basic flow and two-dimensional topography, Royal Meteorological Society, Quarterly Journal, 97(411):30-60.

Vergeiner, I., and Lilly, D. K., 1970, The dynamic structure of lee wave flow as obtained from balloon and airplane observations, Monthly Weather Review, 98(3):220-232.

Virgo, S. E., 1966, Hazards of the foehn wind in Switzerland, Weather, 21(9):306-307.

Wexler, A., and Brombacher, W. G., 1951, Methods of measuring humidity and testing hygrometers, National Bureau of Standards Circular No. 512, United States Department of Commerce Washington.

Wild, H., 1901, Ueber den Foehn und Vorschlag zur Beschraenkung seines Begriffs, Denkschriften der Schweizerischen Naturforschenden Gesellschaft, 38:1-215.

Williams, D. T., 1948, A surface micro-study of squall-line thunderstorms, Monthly Weather Review, 76(11):239-246.

Yih, C., 1965, Dynamics of Nonhomogeneous Fluids, MacMillan Co., New York.

Yoshino, M. M., 1971, Die Bora in Jugoslawien: Eine synoptisch-klimatologische Betrachtung, Annalen der Meteorologie (Neue Folge), 5:117-121.

APPENDICES

APPENDIX A

The

30 November-1 December 1970

Wind Storm

(genetically cold)

On the morning of 30 November 1970 a strong surface inversion existed east of the Continental Divide, as indicated by the Denver sounding and the lapse rate along the slopes (Fig. A.1, right hand side). At about 11:00 MST (18 GMT) a cold front passed Boulder/ Denver; this is reflected in the brief increase in surface wind speed (Fig. A.1, left hand side). The dramatic change in the Denver sounding between 30 November 12 GMT and 1 December 00 GMT (5:00 - 17:00 MST) shows how extremely cold the air behind the front was. At the mountain stations the cold front passage is thus indicated by a decrease in temperature but because of the surface inversion, the temperature increases markedly in Boulder. The case is therefore an excellent example of a cold air foehn. (Note the remarkably constant temperature at the mountain and Boulder stations throughout the wind storm.)

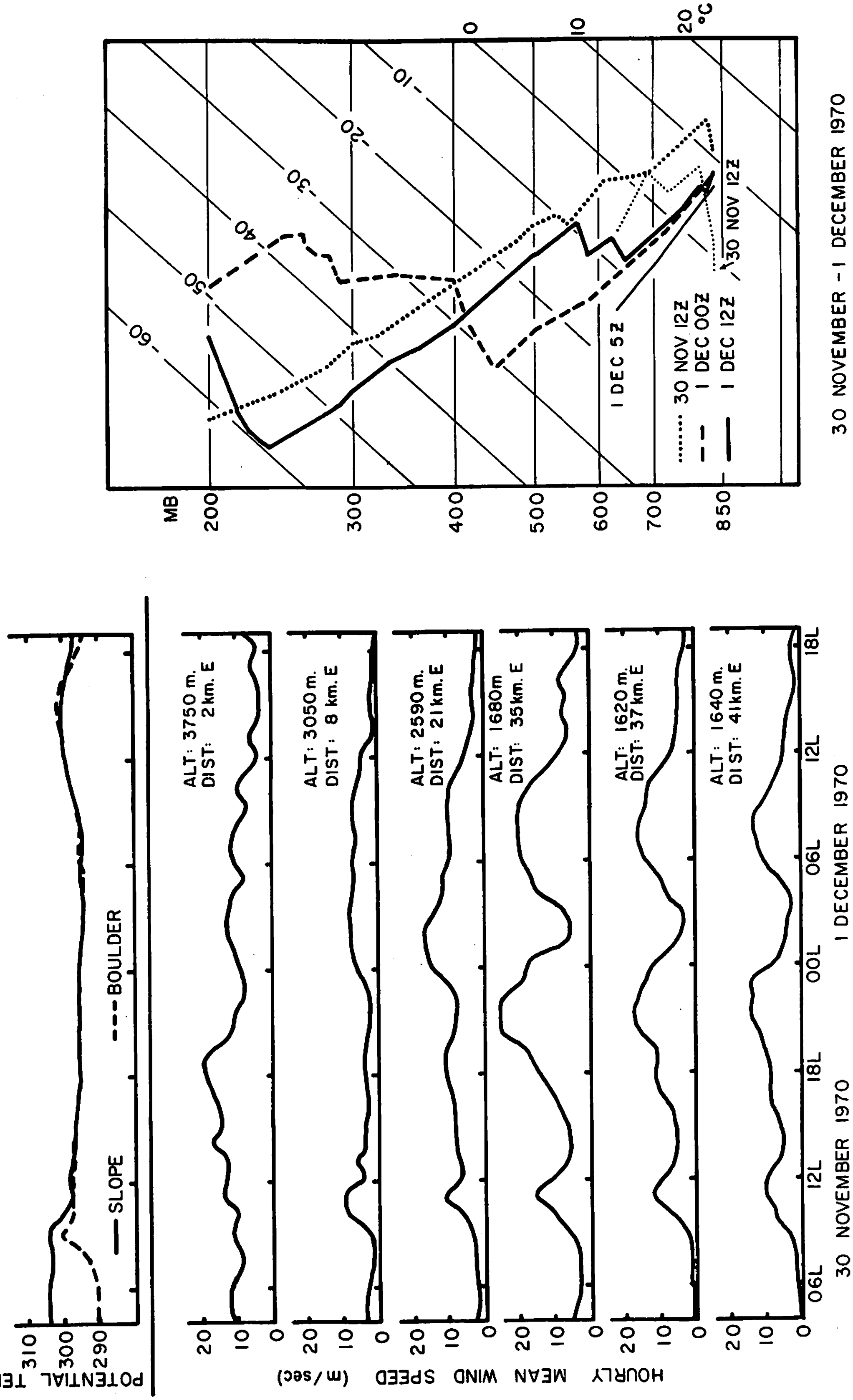
At 17:00 MST (00 GMT) an outstanding inversion, i.e., the top of the cold air, is indicated in the Denver sounding at about the 450 mb level. Shortly afterward the first phase of the wind storm began in Boulder, about 7 hours after the passage of the cold front which at that time was already halfway across Kansas. Between about 1:00 and 3:00 MST a 'pause' in wind speed occurred in Boulder while winds increased at the lowest mountain station. Around 4:00 MST they decreased again at the mountain stations and increased in Boulder. The Denver sounding for 5:00 MST (12 GMT) shows a much lower stable layer as compared with that prior to the first phase of the wind storm. As a matter of fact, two isothermal layers are apparent, one at 650 mb and the other at 590 mb (one might speculate that they may have perhaps been responsible for

Fig. A.1. Temperature and wind speed - 30 November-1 December 1970 storm (from Brinkmann and Zipser 1972).

Right: Denver soundings (heavy lines) and temperatures along the slopes (thin lines).

Left top: Potential temperature along the slopes (average for stations B-1, C-1 and D-1) and in Boulder.

Left bottom: Hourly mean wind speeds at the stations (from top to bottom) D-1, C-1, Gold Hill, Southern Hills, Kent and 76th Street.



POTENTIAL TEMP (°K)
310
300
290
SLOPE
BOULDER
HOURLY MEAN WIND SPEED (m/sec)
ALT: 3750 m. DIST: 2 km. E
ALT: 3050 m. DIST: 8 km. E
ALT: 2590 m. DIST: 21 km. E
ALT: 1680 m DIST: 35 km. E
ALT: 1620 m. DIST: 37 km. E
ALT: 1640 m. DIST: 41 km. E
06L
12L
18L
00L
06L
12L
18L
30 NOVEMBER 1970
1 DECEMBER 1970
MB
200
300
400
500
600
700
850
0
10
20 °C
-10
-20
-30
-40
-50
-60
30 NOV 12Z
1 DEC 00Z
1 DEC 12Z
1 DEC 5Z
30 NOV 12Z
30 NOVEMBER - 1 DECEMBER 1970

the two wind maxima in Boulder). Winds at the next higher mountain station, however, show hardly any fluctuations except for those caused by the frontal passage. At the highest mountain station winds appear to be completely independent of events in Boulder, not even the frontal passage is detectable.

Fig. A.2 shows the surface and 500 mb maps for 1 December 1970, 5:00 MST (12 GMT), the beginning of the second phase. By this time the surface front had moved into Missouri. The 500 mb flow pattern was marked by a westerly flow with a maximum north of Denver. Wind speeds above Denver at that level were 31 m s^{-1}, with a u-component of 28 m s^{-1} (about one standard deviation above the mean, see Fig. 13.9), while maximum gusts at the Boulder stations exceeded 40 m s^{-1} during the storm (see Table 6.2).

The potential temperature and west wind component cross section between Salt Lake City and Dodge City for the same time, i.e., the beginning of the second wind phase, is marked by a cold air dome west of Denver (Fig. A.3). (The fact that neither are the stations located on a straight east-west line nor exactly along the streamlines must be kept in mind in the following discussion.) The two stable layers at Denver, around 300-302°K and 305-310°K are also evident in the soundings of the other stations. The lower layer slopes downward between Salt Lake City and Dodge City and is thus probably continuous with the surface cold front which passed Dodge City around 17:00 MST (00 GMT), 1 December 1970.

Fig. A.4 shows the hourly station pressure for Eagle, Boulder and Denver for the whole period. Clearly, the cold front passed

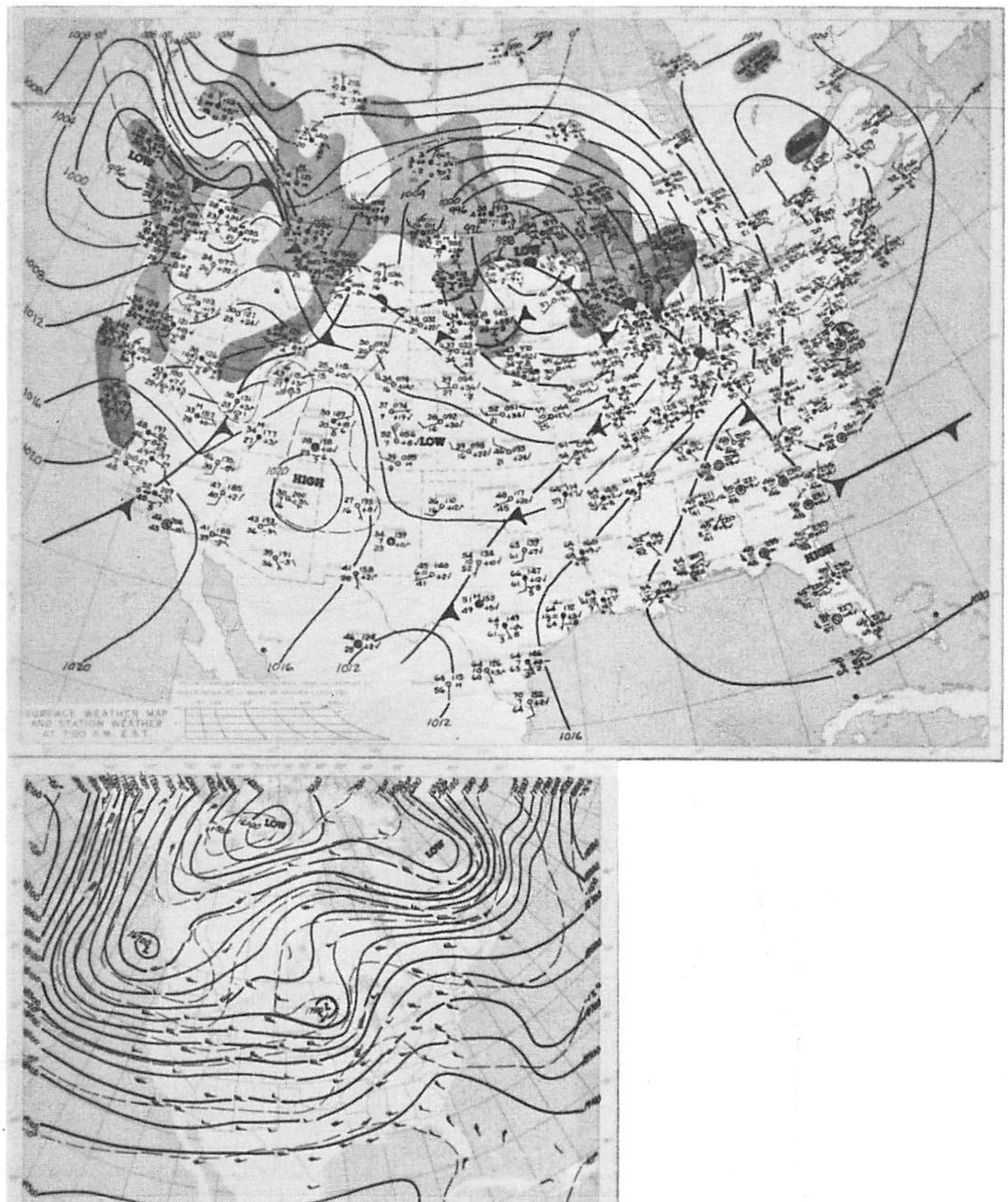


Fig. A.2. Surface weather map (top) and 500 mb height contours (left) - 1 December 1970, 00 GMT.

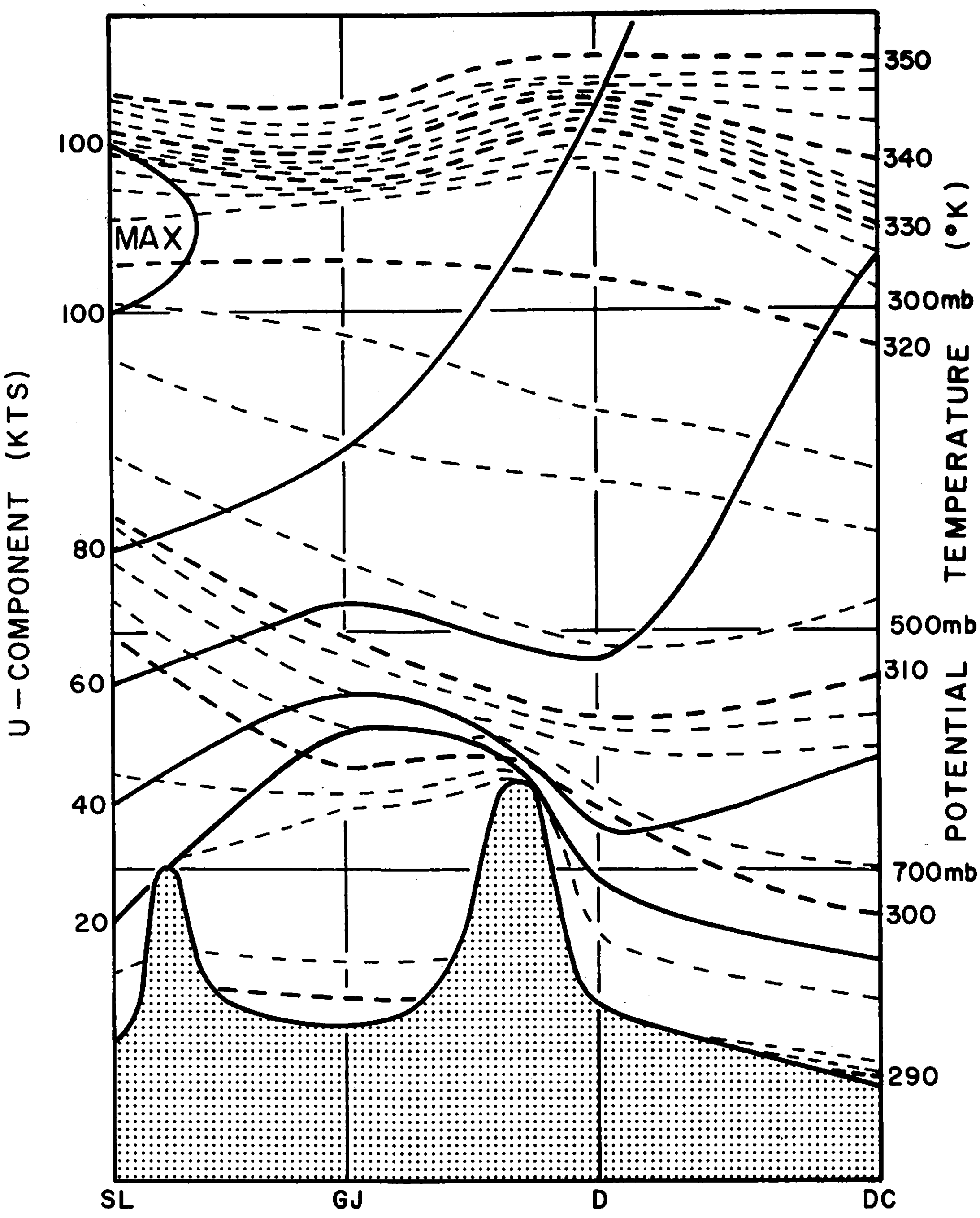


Fig. A.3. Potential temperature and west wind component cross section between Salt Lake City, Utah, and Dodge City, Kansas, based on radiosonde data for Salt Lake City, Grand Junction, Denver and Dodge City, for 1 December 1970, 12 GMT (5 MST). The mean potential temperature along the slopes during the storm was 294°K.

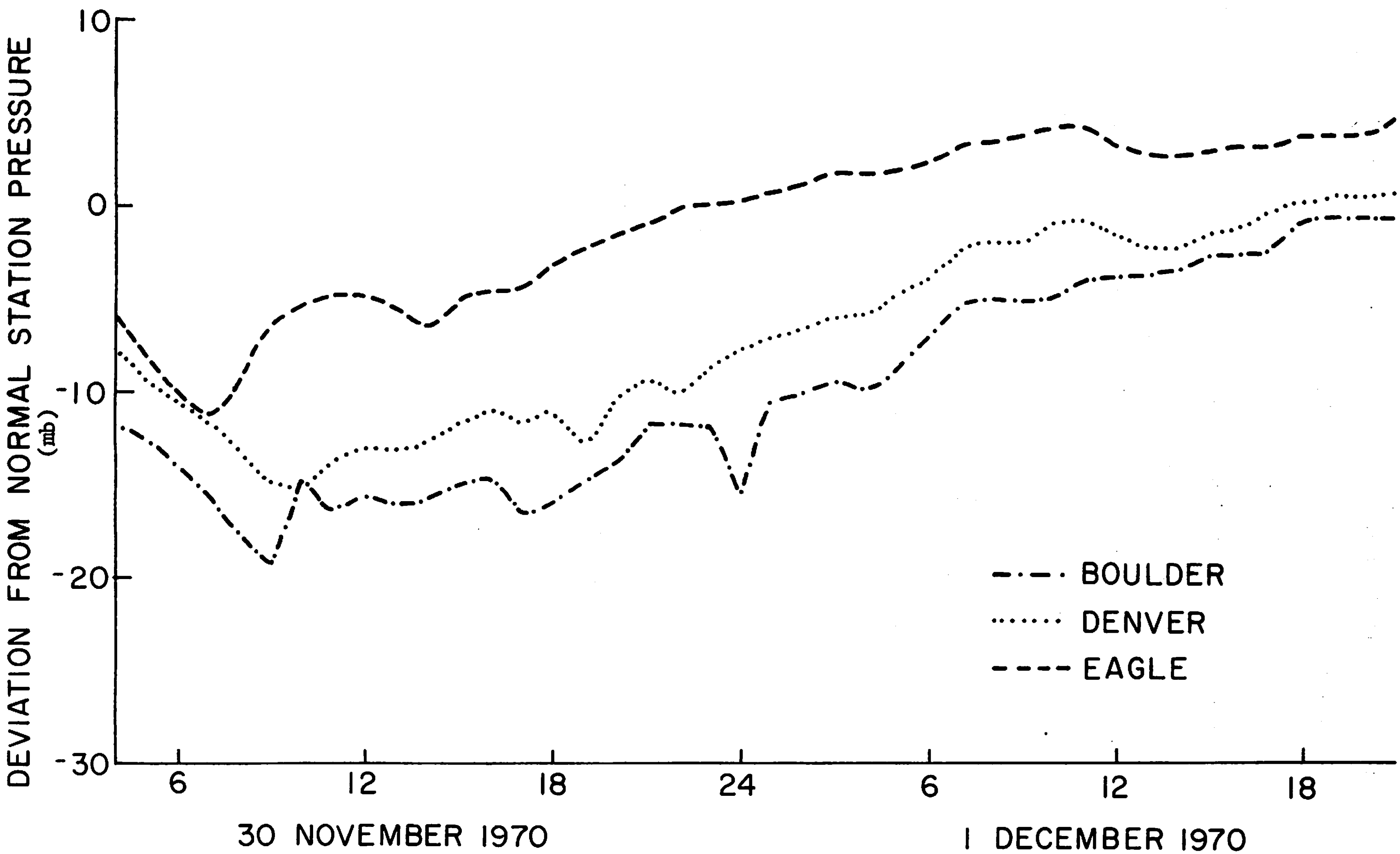


Fig. A.4. Hourly station pressure for Eagle, Boulder and Denver - 30 November-1 December 1970 storm.

Eagle at about 7:00 MST and about three hours later reached the Boulder/Denver area. The pressure rose abruptly at Eagle and less so at Boulder/Denver, creating a large pressure gradient across the mountains. The beginning and end of the first storm phase at the foot of the mountains is indicated by pressure dips in Boulder at 17:00 and 24:00 MST. Small dips, but out-of-phase with Boulder, are also indicated in Denver's pressure data. The beginning and end of the second phase are also marked by pressure dips in Boulder, at 4:00 and 9-10:00 MST, but these are less pronounced. Throughout the whole period, pressure was rising at all stations.

APPENDIX B

The

7 January 1969

Wind Storm

(genetically warm)

This case is not so much a good example of an average warm wind as an example of an outstanding wind storm and shows some of the problems encountered in the analyses discussed in the main text.

On 6 January 1969 a marked temperature rise was observed in the lee of the Divide. This was followed by some winds in Boulder which were, however, not very high and decreased again during the night.

At 5:00 MST (12 GMT) of 7 January 1969 a strong surface inversion existed over Denver, as indicated in the Denver sounding (Fig. B.1, right hand side) which also shows two stable layers, one at 640 mb and another at 405 mb. At that time winds were blowing at the lowest mountain station. Throughout the day high winds continued along the lee slopes and the temperature rose only slightly (Fig. B.1, left hand side).

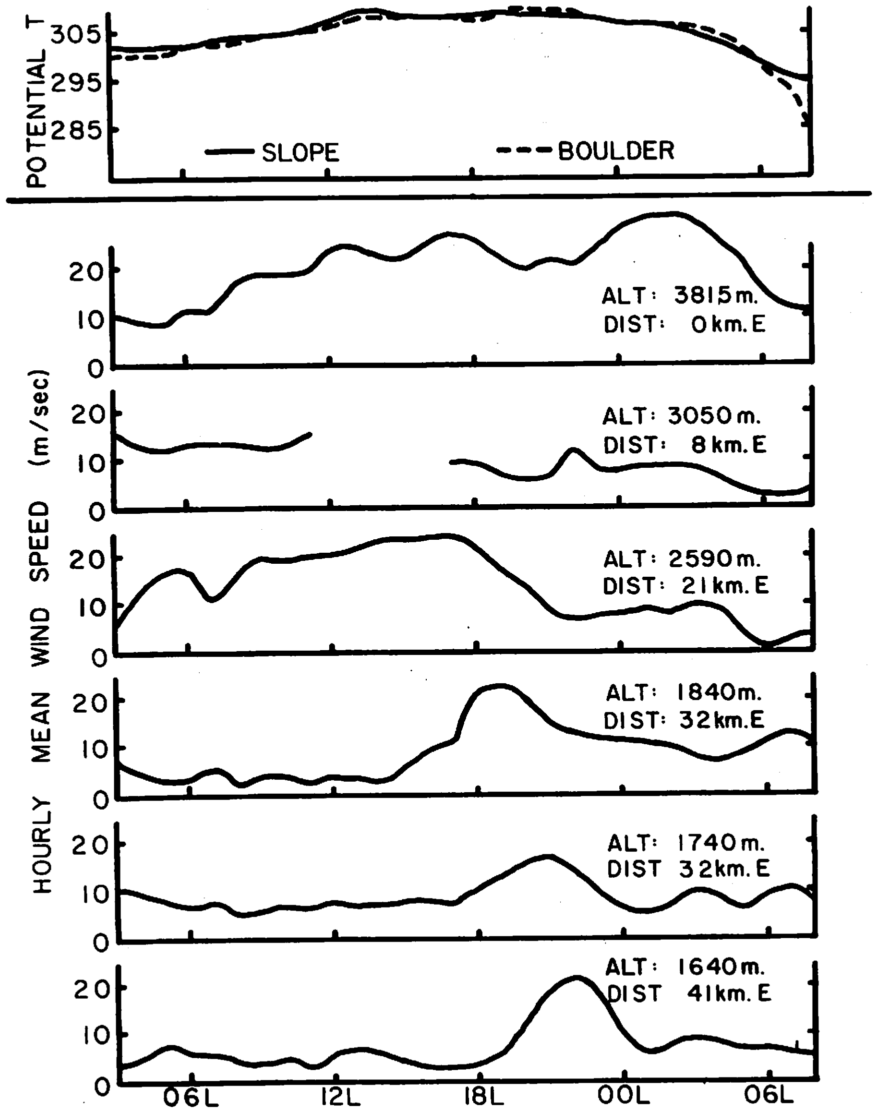
At 17:00 MST (00 GMT) low level warming due to subsidence is indicated in the Denver sounding but this is not reflected in the lee slope and Boulder temperatures. The lowest stable layer has disappeared and the upper one is found at the 500 mb level. The surface map for that time (Fig. B.2) shows a deepening low pressure system to the north, and at the 500 mb level winds were strong and almost due westerly over Colorado. Above Denver wind speeds at that level were 41 m s^{-1}, with a u-component of 39 m s^{-1}, while maximum gusts at the Boulder stations exceeded 45 m s^{-1} during the storm (see Table 6.2). In Fig. B.3 the west wind component cross section for the same time is very similar to that for the 1 December

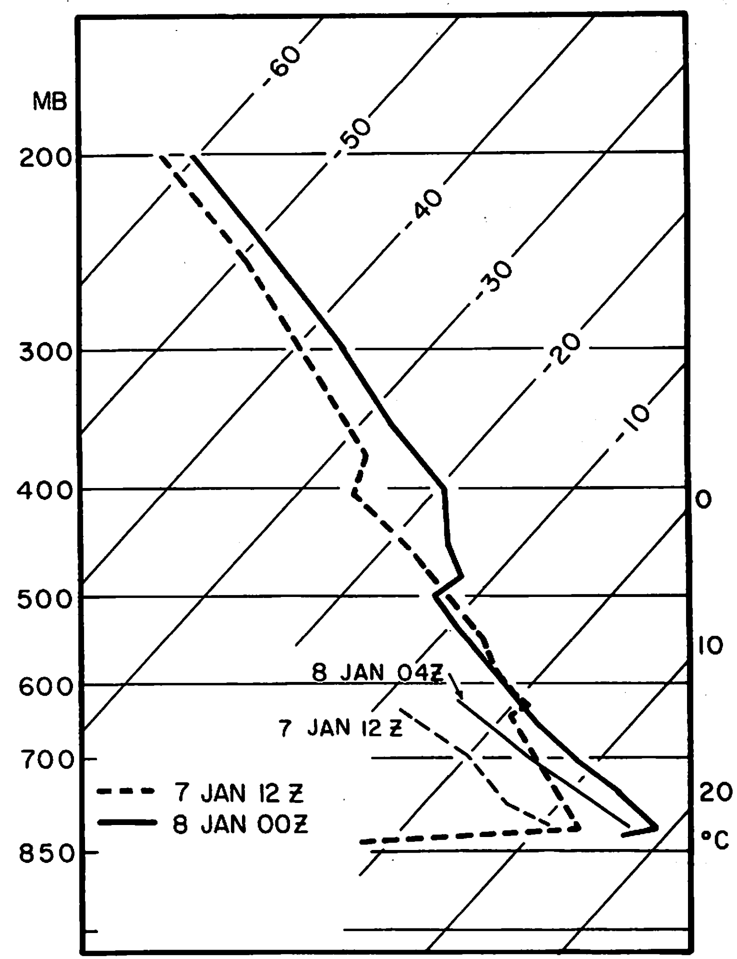
Fig. B.1. Temperature and wind speed - 7 January 1969 storm (from Brinkmann and Zipser 1972).

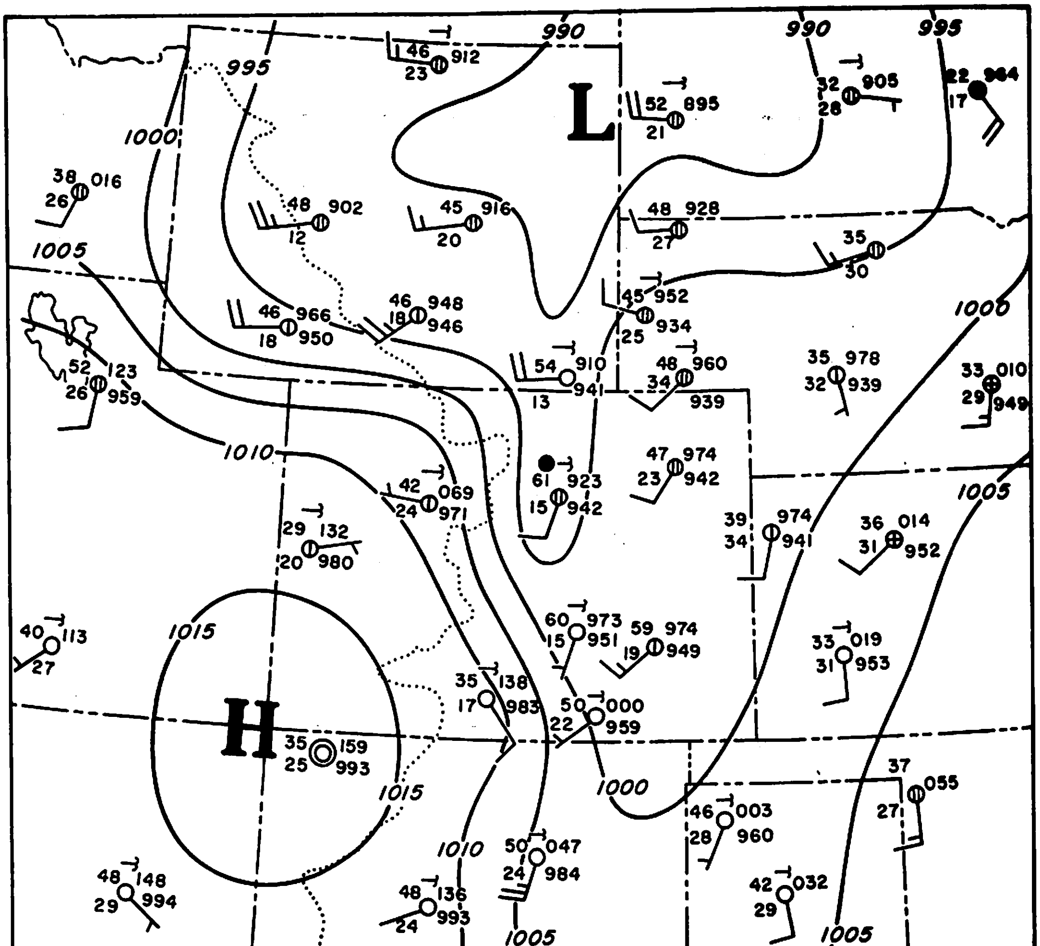
Right: Denver soundings (heavy lines) and temperatures along the slopes (thin lines).

Left top: Potential temperature along the slopes (averages for stations B-1, C-1 and D-1) and in Boulder.

Left bottom: Hourly mean speeds at the stations (from top to bottom) Mines Peak, C-1, Gold Hill, NCAR, Boulder Camera, and 76th Street.


POTENTIAL T
285
295
305
SLOPE
BOULDER
HOURLY MEAN WIND SPEED (m/sec)
ALT: 3815 m.
DIST: 0 km. E
ALT: 3050 m.
DIST: 8 km. E
ALT: 2590 m.
DIST: 21 km. E
ALT: 1840 m.
DIST: 32 km. E
ALT: 1740 m.
DIST 32 km. E
ALT: 1640 m.
DIST 41 km. E
06L
12L
18L
00L
06L


MB
200
300
400
500
600
700
850
-60
-50
-40
-30
-20
-10
0
10
20
°C
8 JAN 04Z
7 JAN 12 Z
7 JAN 12 Z
8 JAN 00Z



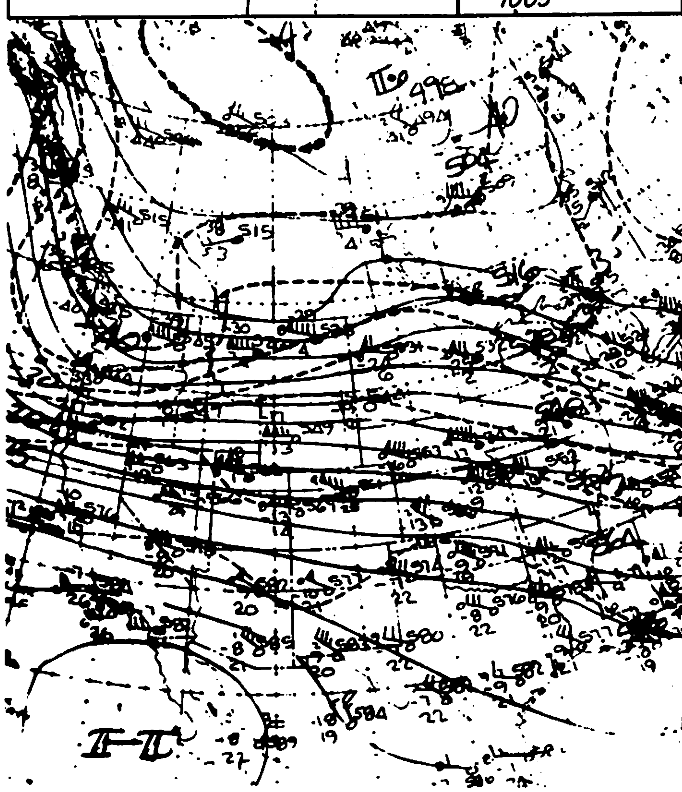


Fig. B.2.

Surface weather map (top; from Julian and Julian 1969) and 500 mb height contours (bottom) - 8 January 1969, 00 GMT.

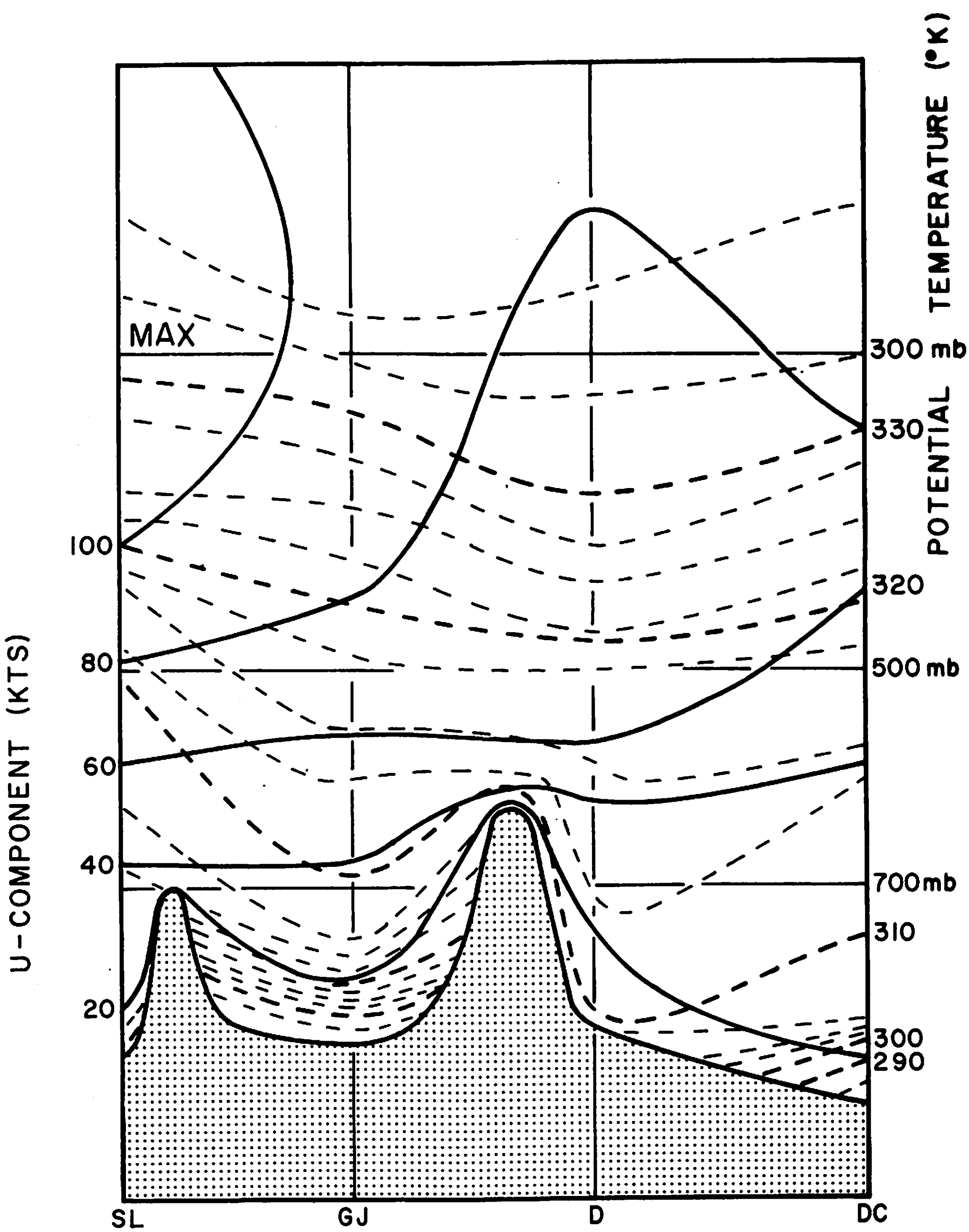


Fig. B.3. Potential temperature and west wind component cross section between Salt Lake City, Utah, and Dodge City, Kansas, based on radiosonde data for Salt Lake City, Grand Junction, Denver and Dodge City, for 8 January 1969, 00 GMT (17 MST). The mean potential temperature along the slopes during the storm was 308°K.

1970 cold case, with rapid increases and high winds at low levels and relatively low speeds in the upper atmosphere. The stable layer over Denver at 500 mb can also be seen over Salt Lake at a slightly higher level.

Within an hour of these observations wind speeds began to decrease at the lowest mountain station (but continued to be high at the Divide) and increased in Boulder (note the obvious movement of the wind maximum down the slopes, i.e., between the 3rd and 2nd station, and out into the plains, i.e., between the 2nd and 1st station). The storm lasted only about 3 hours in Boulder and by midnight winds had decreased everywhere except at the Divide, but the cold front did not pass through the area until the next day. Some very critical air flow conditions were met for a very short period of time.

Fig. B.4 shows the hourly station pressure for Eagle, Boulder and Denver for the whole period. The pressure began to decrease at all stations at about the same time, approx. 21:00 MST. The decrease was, however, more rapid in the lee as compared with the windward side, resulting in an increasing pressure gradient across the mountains. The onset of the wind storm in Boulder is marked by a rapid drop in pressure. Throughout the storm period the pressure continued to decrease and, after a final small dip, increased again as the wind in Boulder died down. Throughout the storm period the pressure decreased at all stations.

The major question that arises from the above description of one of the most outstanding and damaging storms is the level of the stable layer. The upper one first at 400 mb and then at 500 mb appears to be too high since strong winds were blowing in the mountains throughout the day. Perhaps the lower layer is the correct one but it was gone by 17:00 MST when winds were still strong in the mountains. Furthermore, the upper stable layer at 500 mb is quite high considering that a wind storm occurred in Boulder within one hour.

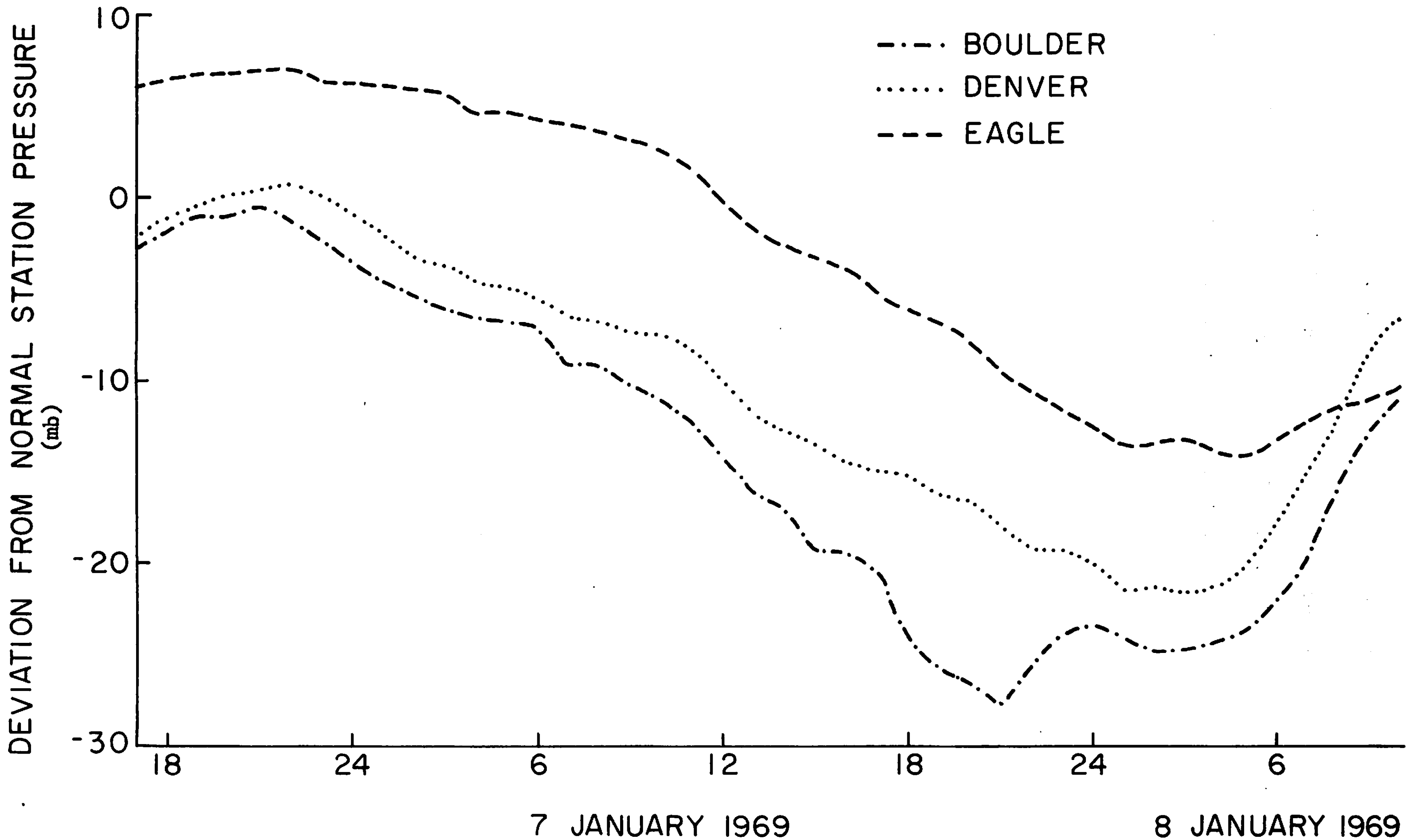


Fig. B.4. Hourly station pressure for Eagle, Boulder and Denver - 7 January 1969 storm.

University of Colorado
Institute of Arctic and Alpine Research
Boulder, Colorado 80302