



STUDIES OF THE LOWER ATMOSPHERE WITH A VHF WIND PROFILER

By

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Abstract

The theory and application of VHF wind profiler observations of the lower atmosphere (below 20 km altitude) are discussed. Beginning with a review of conventional atmospheric sounding systems and time series analysis as applied to wind profilers, we move on to a detailed investigation of some technical aspects: the effects of phasing errors in the antenna array subsystems; a comparison of winds measured with the spaced antenna technique and windfinding radiosondes; and a critical examination of balloon ascent rates as a measure of atmospheric vertical velocity.

The altitude coverage of wind profilers is of major interest to those who wish to make use of these instruments on a regular basis, and we present a quantitative, predictive model of profiler performance based on historical or climatological thermodynamic data. We evaluate the model by comparison between two independent profiler systems and nearby radiosonde station records, and make some speculative predictions over a wide range of latitudes.

Continuing the theme of probing atmospheric vertical structure, we evaluate an algorithm for tropopause altitude retrieval from profiler returns. This is complemented by the description and evaluation of a cluster-mean approach for the location and tracking of multiple stable layers. The possibilities of deriving temperature and humidity profiles from profiler returns are discussed.

The remainder of the thesis is devoted to another major strength of the VHF wind profiling technique: the ability to make continual, real-time observations over an extended period. The applications to regional forecasting are highlighted through two case studies. Firstly, we present a detailed examination of an intense midlatitude cyclone through a combination of profiler, radiosonde, and regional mesoscale

assimilations. The intercomparison between the three techniques allows us to confirm the applicability of profiler returns; the unique insights provided by the profiler are highlighted. The second case study details a relatively common occurrence around southern Australia: the prefrontal trough. This type of system is a major source of forecasting error, and we draw some conclusions regarding its evolution.

Taken as a whole, this thesis represents a detailed examination of wind profiler applications to studies of lower atmosphere structure and mesoscale meteorology. It is a timely investigation and review of the field, given the rapidly increasing interest being shown by meteorological organisations with regards to radar wind profilers.

Acknowledgements

We don't do it for the glory

We don't do it for the recognition

We do it because it needs to be done.

Because if we don't, no one else will.

And we do it even if no one knows what we've done.

Even if no one knows we exist. Even if no one knows we ever existed.

— David Morrell, *First Blood*

An undertaking the size of a research doctorate cannot be conducted in isolation—although late at night, staring at a computer terminal, with a few hundred examples of the Dynamic Ribbon Device on cans in the background, one could be forgiven for thinking otherwise.

This work was carried out under the supervision of Dr. Bob Vincent, to whom I express my heartfelt appreciation. His scientific insight, academic and moral support, and willingness to chat about any and all aspects of the Ph.D. experience were invaluable. Dr. Iain Reid is also deserving of thanks, for his all-round assistance and advice, and for assuming the supervisory role during Bob's absence. Dr. Graham Elford and the late Dr. Basil Briggs also provided ample opportunity for discussion—together, they reinforced my regard for the ways of the School of Classical Science. I should also like to thank Dr. Peter May of the BMRC and Dr. Steve Eckermann of NRL, both past students of the Atmospheric Group at Adelaide, for discussions over uncounted beers regarding Life, the Universe and Everything. Peter supplied the Darwin profiler data used in Chapter 4, and I thank him for that in particular. The other members

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The SAFEx campaign was conducted in collaboration with The Centre for Dynamical Meteorology and Oceanography at Monash University (Clayton): directly, with Dr. Michael Reeder and soon-to-be-Dr. Morwenna Griffiths. Their aid and assistance, both during their visits to Adelaide for the observational campaigns, and on the occasions of my visits to Monash for collaborative analysis of the SAFEx dataset, were invaluable. The three months I spent at Monash was without doubt the most enjoyable period of my doctoral studies, due in no small part to their efforts. Chapter 6 in particular is a part-result of our collaborative research, and it would be remiss of me not to acknowledge their contributions. Michael’s suggestion of the instant occlusion comparison in the same Chapter should also be acknowledged, together with initial discussions of the prefrontal trough work in Chapter 7. I credit the majority of my knowledge of meteorology and “PV-thinking” to osmotic processes whilst around Michael, although any misapprehensions are purely my own. Morwenna was responsible for extracting the RASP and McIDAS analyses and converting them to an IDL-readable format, and had a major part in formulating the synoptic and mesoscale background to Chapter 6. The assistance she provided in introducing me to the Department of Materials Engineering at Monash as part of the investigation of balloon ascent rate fluctuations, made a significant portion of that work possible. Finally, by acting as a sounding-board on all aspects of the SAFEx work for a number of years (despite never convincing me to write “SAFEX”), the research process was made immeasurably more fruitful.

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Through me the Road to the City of Desolation;
Through me the Road to Sorrows Diurnal;
Through me the Road to the Lost of Creation.

Justice moved my Great Maker; God Eternal
Wrought me; the Power, the unsearchably
High Wisdom, and the Primal Love supernal.

Nothing ere I was made was made to be,
Save things eterne, and I eterne abide.
Lay down all hope, you that go in by me.

—*The Divine Comedy of Dante Alligheri the Florentine,*
First Canticle, Third Cantos.

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Chapter 1

INTRODUCTION

So you wish to stride into the infinite?

Then explore first the finite in all directions.

— *Goethe*

It is human nature to explore, to question the environment, to seek answers; "*To boldly go...*", in the words of Gene Roddenberry. Yet even today, more than a quarter century after Man walked on the Moon, our knowledge of the terrestrial sphere is essentially limited to the surface. The depths of the oceans remain hidden; the interior of the Earth is a mystery; and the vagaries of the atmosphere defy analysis. We can predict the course of comets over centuries with confidence, but next week's weather is accepted as a great unknown.

Of course, it has always been easier to look outwards rather than inwards. The study of outer space is made easier because the Earth itself gives us a platform for observations, and the ability to base sensors (from our own eyes to radio telescopes and beyond) which can examine phenomena at a distance. Only in the last few decades have we been able to apply the same techniques of "remote sensing" to the the Earth as a whole, using artificial satellites that can map the surface and profile the atmosphere. These measurements have, for the first time, enabled us to investigate climate—*global* weather—and changes in the environment on large scales.

But what about the smaller scales, and those things that affect our lives on a day to day basis? Satellite observations do not provide us with the temporal or spatial

resolution required to investigate the detail in, or the evolution of, mesoscale systems such as frontal passages; much less the small-scale dynamics associated with such events. These are the areas upon which this thesis concentrates: observing scales in the atmosphere smaller than those generally observed by satellites, with a better time-resolution than that obtainable by conventional balloon-borne sounding systems.

With this in mind, it is worth beginning with an exposition of scales and structure in the atmosphere. This is followed by a review of atmospheric observation systems, particularly those utilised in this thesis. A summary of the scope of the remainder of this work completes the introduction.

1.1 The Atmosphere

1.1.1 Vertical Structure

As recently as 1900, it was possible for a reputable scientist to conclude, on the basis of the limited data of the time and deductions from radiation theory, that temperature must continue to decrease with altitude to the outer edge of the atmosphere. But even as such conclusions were being reached, a series of observations were being carried out that culminated in the announcement of the existence of an “isothermal layer” near 11km altitude [*Teisserenc de Bort*, 1902].

Nearly a century, and a huge number of observations later, considerable detail is available about the vertical structure of the atmosphere. Although there are a number of ways of classifying regions in the vertical, the most common—and most appropriate as far as the work presented herein is concerned—is by temperature variation. Figure 1.1 shows a temperature profile for the Earth’s atmosphere taken from a mid-latitude equinox model. The regions are characterised by the change in temperature with height; in ascending order, they are the troposphere, stratosphere, mesosphere and thermosphere. At higher altitudes, the conventional concept of temperature breaks down as the mean free path of the constituent particles increases. More generally, we could talk about the “lower” (up to about 10 km altitude), “middle” (between 10 km and 100 km), and “upper” (above 100 km) atmosphere.

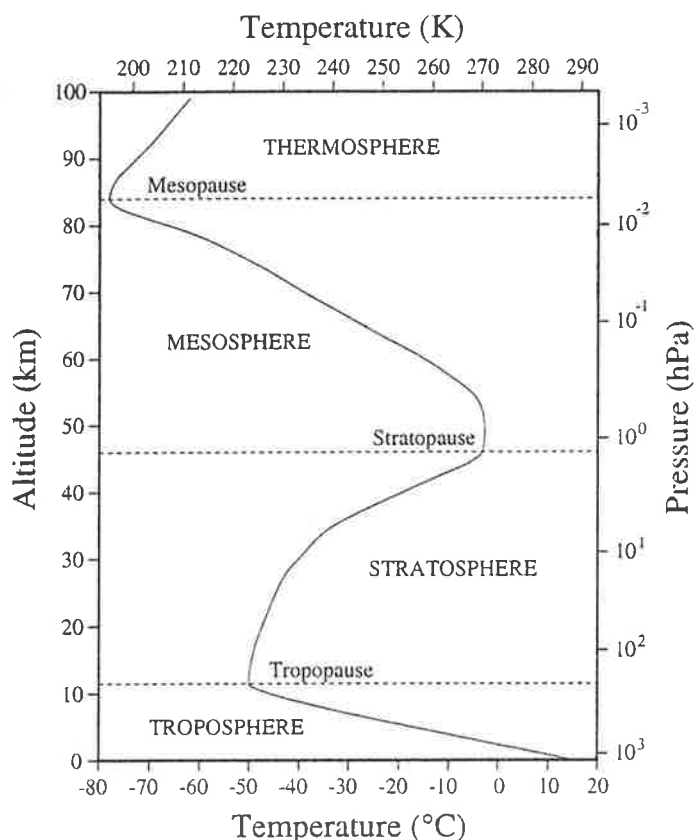


Figure 1.1: The Earth's atmospheric temperature profile and associated nomenclature, based on the U.S. Standard Atmosphere (1976).

In this thesis, we concentrate exclusively on the troposphere and stratosphere, together with the intervening transition region known as the tropopause (Teisserenc de Bort's "isothermal layer"). Weather systems have little direct effect above the troposphere, although the atmospheric wave motions generated by such systems can and do propagate throughout the atmosphere (a result of energy conservation, together with the decrease of pressure with altitude). The observation of weather-driven effects in the stratosphere can be carried out by a number of complementary techniques, not all of which are applicable higher in the atmosphere.

The troposphere is characterised by an approximately constant decrease of temperature with height, due to the absorption and re-radiation of incoming solar energy at

the ground; the model lapse rate in Figure 1.1 is about 6.5 K km^{-1} . Tropospheric temperatures are governed by radiative and convective exchange, and standard radiative-convective models of the lower atmosphere (*e.g.* Chamberlain & Hunten [1987]) reproduce the main features of the tropospheric temperature distribution, even under uniformly absorbing (“grey”) conditions. There are some problems, however: for example, a grey atmosphere heated from below should produce a warmer (*i.e.* lower) tropopause in the tropics than near the poles, simply because the ground temperature is warmer. In fact, the tropical tropopause lies at about 16 km with a temperature of about 195 K, whereas over the poles it is near 8 km (about 225 K): see Holton [1992] for a discussion of the breakdown of the greyness approximation.

Temperature in the stratosphere increases with height, due to the ability of ozone (O_3) molecules to absorb both incoming solar ultraviolet, and outgoing terrestrial infrared radiation. The distribution of ozone varies markedly over both altitude and latitude, with the peak tending towards lower altitudes in the tropics and higher abundances near the poles due to the meridional circulation. This affects the temperature profiles—the tropical tropopause is sharply defined, with an abrupt temperature inversion and rapid rise in the lower stratosphere, whereas in polar regions the lower stratosphere is nearly isothermal.

Of course, the description above is quite simplistic. Detailed model calculations include many other contributing effects, one example being the variation of convection with latitude. Convection is driven more strongly in the tropics because solar radiation is more abundant, and because the high water vapour content there increases the cooling of the upper troposphere (at the same time inhibiting the direct cooling of the ground by radiation to the stratosphere). The interplay of such effects requires elaborate modelling, but the complex feedbacks that result can obscure the physics at work. The subject is covered in more detail by a number of authors; the reader is referred to Craig [1965], Holton [1992] and Andrews *et al.* [1987] as examples.

1.1.2 Scales

Visual observations provide the first measures of scale in the atmosphere—clouds evolve over about half an hour, weather changes on a time scale of a day or two (at least in middle and high latitudes), and the seasons have an annual cycle. Spatial scales are more difficult to estimate without reference to some other source of information: we could compare cumulus clouds to mountains, or tornados to buildings, for example. The first real success regarding linked spatial and temporal scales did not occur until the turn of the century, when the Norwegians made use of the spacing of European cities to set up an observing array which allowed the investigation of cyclones and fronts at the surface. Today, an examination of tropospheric “weather maps” indicates that there are characteristic spatial scales (the spacing of high and low pressure systems, for example). We refer to this as the *synoptic* scale, and it primarily concerns features such as cyclones and anti-cyclones with spatial scales of 500–2000 km and temporal scales of 3–15 days. Careful study reveals an even larger *planetary* scale (8000–40000 km, 15+ days), upon which these pressure patterns are superimposed (in the stratosphere, these planetary scales are the dominant feature). Both the planetary and synoptic scales make use of observing networks, with sensors spaced at intervals of around 100 km, to form pictures of the environment.

At the other end of the spectrum is the *molecular* scale on which, ultimately, all atmospheric exchanges occur. The smallest scale of interest to meteorologists, though, is the *micrometeorological* scale. This encompasses length scales of a metre or less, time scales of seconds, and is concerned with such phenomena as surface roughness, heat fluxes, and turbulence in the boundary layer (the lowest few hundred metres of the atmosphere, where frictional effects are still important). A single observing station at the surface is usually sufficient to observe these effects. Cloud physics is another example of microscale meteorology.

In the last few decades, it has become apparent that many of the local weather effects exist between these scales. As foreshadowed by *Ligda* [1951], a radar meteorologist:

It is anticipated that radar will provide useful information concerning the

structure and behaviour of that portion of the atmosphere which is not covered by either micro- or synoptic-meteorological studies. We have already observed with radar that precipitation formulations which are undoubtedly of significance occur on a scale too gross to be observed from a single station, yet too small to appear even on sectional synoptic charts. Phenomena of this size might well be designated as mesometeorological.

Today we designate them as *mesoscale*, which is far less of a mouthful! It is interesting to note that the term was coined to describe scales that could *not* be observed by contemporary systems (5–500 km, with timescales of a few hours to days), rather than to identify a particular observation. Such a wide range of phenomena fall into the mesoscale—thunderstorms, rainbands, tropical cyclones, and cold-frontal circulation, for example—that it is often subdivided according to horizontal dimensions into the meso- α and meso- β (and even further into the Greek alphabet by some schemes). This is illustrated in Figure 1.2, which compares a number of meteorological events and their associated horizontal scale according to a number of schemes.

Of course, it is a mistake to think of the atmosphere as quantised into distinct scales. The spectrum of energy in the atmosphere is continuous, and although it displays distinct peaks at certain frequencies corresponding roughly to the scales outlined above [Vinnichenko, 1970], energy transfer between scales is ubiquitous. This is known as *scale interaction*, and is most often seen in the influence of the mean flow on a disturbance (although a large enough disturbance can affect the mean flow, leading to secondary instabilities). One example is the effect of gravity (buoyancy) waves on convection—gravity waves can be generated by convective instabilities, but may also serve to organise convection in their own right.

A more detailed discussion of scales in the atmosphere is presented by Ray [1986].

1.1.3 Stability

Much of the work presented herein is concerned with motion in the atmosphere. It is therefore appropriate to introduce some of the background definitions and terminology which are associated with such motions. In particular, we shall make frequent use of

Disturbances	Scale	Ligda (1951)	Orlanski (1975)	Fujita (1981)
	10,000 km	SYNOPTIC	MACRO	α
	1,000 km		α	MASO
	100 km	MESO	MESO β	α
	10 km		γ	β
	1 km	MICRO	α	α
	100 m		MICRO β	MISO
	10 m		γ	α
	1 m			MOSO
				β
				MUSO

Figure 1.2: A comparison of horizontal scales in the atmosphere, according to schemes proposed by *Ligda* [1951], *Orlanski* [1975] and *Fujita* [1981] (after *Fujita* [1986]).

the concept of stability in the atmosphere.

Consider the case of an ideal gas undergoing an adiabatic process (that is, one in which no heat is exchanged with the surroundings). Then, the first law of thermodynamics may be expressed as:

$$c_{pd} d(\ln T) - R_d d(\ln p) = 0 \quad (1.1)$$

where T and p are respectively temperature and pressure, $c_{pd} = 1005.7 \pm 2.5 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat of dry air at constant pressure, and $R_d = 287.04 \text{ J kg}^{-1} \text{ K}^{-1}$ is the gas constant in dry air. If we integrate from a state at pressure p and temperature T to a state at pressure p_s (usually 1000 hPa) and temperature θ , and introduce $\kappa_d = R_d/c_{pd} = 0.2854 \pm 0.0007$, then:

$$\theta = T \left(\frac{p_s}{p} \right)^{\kappa_d} \quad (1.2)$$

This defines the *potential temperature* θ as the temperature which a parcel of dry air at pressure p and temperature T would have if it were compressed or expanded

adiabatically to pressure p_s . Potential temperature is conserved for (dry) adiabatic flow, and hence is superior to T as a field variable for characterising the thermodynamic state of the atmosphere.

Taking the logarithm of Equation 1.2, and applying the hydrostatic equation:

$$dp = -\rho g dz \quad (1.3)$$

and the ideal gas equation:

$$p = \rho R T \quad (1.4)$$

yields:

$$\frac{T}{\theta} \frac{\partial \theta}{\partial z} = \frac{dT}{dz} + \frac{g}{c_{pd}} \quad (1.5)$$

We define the (environmental) lapse rate to be:

$$\Gamma = -\frac{dT}{dz} \quad (1.6)$$

so that the dry adiabatic lapse rate Γ_d is the lapse rate of temperature in an atmosphere where θ is constant with height:

$$\Gamma_d = \frac{g}{c_{pd}} \quad (1.7)$$

If $\Gamma < \Gamma_d$, so that θ increases with height, a parcel of air which undergoes an adiabatic displacement from its equilibrium level will be positively buoyant when displaced vertically downwards, and negatively buoyant when displaced vertically upwards. Hence, the parcel will tend to return to its equilibrium level, meaning that the atmosphere is statically stable (or, alternatively, stably stratified). From the temperature profile of Figure 1.1, we expect that the static stability of the stratosphere should be much greater than that of the troposphere.

The most convenient measure of stability is the buoyancy frequency, N (also known as the Brunt-Väisälä frequency), defined by:

$$N^2 = g \frac{\partial \ln \theta}{\partial z} \quad (1.8)$$

so that:

$$N^2 = \frac{g}{T} \left(\frac{dT}{dz} + \Gamma_d \right) \quad (1.9)$$

where N gives the frequency of adiabatic oscillations for a fluid parcel displaced vertically from its equilibrium level in a stably stratified atmosphere (see, for example, *Holton* [1992]).

The treatment above has neglected moist effects, which is a reasonable assumption in the stratosphere and upper troposphere. Indeed, potential temperature θ is a useful, conserved quantity provided we are interested in dry processes. The corresponding conserved variable for moist processes is the *equivalent* potential temperature, θ_e , defined as the potential temperature that a parcel of air would have if all its moisture were condensed out and the resultant latent heat used to warm the parcel [*Holton*, 1992]. A full treatment of θ_e is given by *Bolton* [1980], following the work of *Simpson* [1978]. *Bolton* [1980] carries out a thermodynamic integration by raising a parcel of air dry adiabatically to its lifting condensation level, then pseudo-wet adiabatically to a great height (dropping out condensed water as it is formed), and finally bringing it down dry adiabatically to a standard level of 1000 hPa: the final temperature is θ_e , for which the computational expression is:

$$\theta_e = T \left(\frac{1000}{p} \right)^{\kappa_d(1-0.28 \times 10^{-3}q)} \times \exp \left[\left(\frac{3.376}{T_L} - 0.00254 \right) \times q(1 + 0.81 \times 10^{-3}q) \right] \quad (1.10)$$

where T , p , and q are the absolute temperature (K), pressure (hPa) and mixing ratio (g kg^{-1}) at the initial level; T_L is the absolute temperature at the lifting condensation level, given by any of:

$$T_L = \begin{cases} \frac{1}{\frac{1}{T_D - 56} + \frac{\ln(T/T_D)}{300}} + 56 & \text{(dewpoint)} \\ \frac{1}{\frac{1}{T - 55} - \frac{\ln(U/100)}{2840}} + 55 & \text{(humidity)} \\ \frac{2840}{3.5 \ln T - \ln e - 4.805} + 55 & \text{(mixing ratio)} \end{cases} \quad (1.11)$$

where T_D is the (absolute) dew point temperature, U is the percentage relative humidity, and e is the water vapour pressure given by:

$$e = \frac{pq}{622 + q} \quad (1.12)$$

Bolton [1980] quotes maximum errors in the derived θ_e values of about 0.3 K, mainly due to neglecting the variation of c_{pd} with temperature and pressure: note that this simplification also affects the value of θ calculated by Equation 1.2.

1.2 Upper-Air Observing Systems

The measurement of meteorological quantities at or near the surface is straightforward, provided one observes certain conventions regarding the location of the sensors away from obstructions and local effects such as drainage currents and the like. Upper-air measurements are more difficult, simply because of the remote nature of the observation. We are faced with a choice between placing a sensor within the environment (an *in situ*) measurement, or observing the effect that the environment has on the passage of radiation (through scatter, absorption, refraction, or Doppler-shifting, for example). The latter is labelled *remote sensing*, and can be either active (transmitting energy, and observing what happens to it) or passive (“listening” to natural radiation).

If we wish to evaluate an observing system, there are some obvious criteria: accuracy, resolution, sensitivity, and reproducibility (precision). If we are observing the atmosphere, there is another consideration: representativeness. That is, is our measurement an appropriate sample of the state of the atmosphere? A sensor may be as accurate as we please, but if it is sampling an atypical region then its usefulness is significantly degraded.

1.2.1 Balloon-borne radiosondes

The first airborne measurements were made in 1749, using a thermometer attached to a kite. These were followed by meteographs carried on balloons (although retrieval of the data could take days!) and aircraft, until the launch of the first telemetering radiosonde in 1927. By 1940, small balloon-borne radiosondes were the standard measuring system for profiles of temperature, pressure and humidity. Today’s radiosondes are functionally the same as those of fifty years ago, although technological improvements have been made to the thermodynamic sensors. Typical errors would be 1 hPa in pressure, 0.5 K in temperature, and 5% in relative humidity, up to a maximum altitude of 25 km or higher.

The measurement of winds has undergone a more involved evolution, beginning with tracking of the balloon itself using single or multiple theodolites. The balloon is assumed to travel with the mean wind, which can be calculated by the rate of change

of position of the balloon. Other “angle measurement” methods include tracking of the radiosonde transmissions via radio direction-finding, and radar tracking of reflectors suspended beneath the sonde package. All of these methods, however, become increasingly inaccurate as the balloon’s range increases.

The modern preferred method makes use of VLF signals transmitted by the Omega or Loran-C navigational aid (navaid) beacons. The radiosonde’s VLF receiver retransmits the navaid signals to a base station, where the difference in arrival time from different VLF transmitters is used to locate the sonde in space. This method has significant advantages in vertical resolution and accuracy, which does not suffer greatly as range increases. Errors in the measured wind are estimated at less than 2 ms^{-1} for two-minute averages of ten-second data.

Details about the precise systems used in this thesis are given in Chapter 3.

1.2.2 Satellite observations

Satellite observations of weather began with TIROS-1 in 1960, and have since become an indispensable part of weather observation and forecasting. They provide a unique picture of the Earth’s cloud cover on a vast scale, together with a variety of sounding instruments. Polar-orbiting satellites (such as the TIROS-N/NOAA craft) circle the Earth twice a day at about 850 km altitude, and provide global observations four times a day with approximately 1 km spatial resolution over a scan width of about 2000 km. Geostationary satellites (GOES, GMS, Meteosat) provide continuous observations—normally processed on a half-hourly basis—of a hemisphere from 35000 km above the surface, at about 5 km resolution. However, both systems have relatively poor vertical resolution, in the order of kilometres, for radiance-derived fields such as temperature and moisture. Lower-atmosphere winds can be measured by observing cloud drifts [Hubert, 1979]; absolute vector differences with balloon winds range from 2 to 10 ms^{-1} [Hubert & Whitney, 1971]. More recently, the HRDI (high resolution Doppler imager) and WINDII (wind imaging interferometer) instruments onboard the Upper Atmosphere Research Satellite (UARS) have been used to observe stratospheric and mesospheric winds.

1.2.3 “Weather” radars

In simple terms, a radar transmits a pulse of energy in a selected direction, then “listens” for an echo. Range to the target is computed from the transmission/reception time delay. Although the initial applications of radar were in detecting solid objects, it was soon discovered that hydrometeors (rain, ice, snow, etc.) were also capable of producing a radar return. The conventional weather (precipitation) radar can locate rain and indicate its intensity from the amplitude of the returned signal, using a nearly horizontal beam. Changes in the frequency of the returned signal can be used to determine the radial velocity of the hydrometeors (and hence the wind, assuming that the targets are carried by it) using the Doppler principle. Scanning the look direction of the radar azimuthally allows a picture of the surrounding region to be constructed; if the beam is elevated above the horizon, then a range-to-altitude conversion can be implemented.

Precipitation radars are widely discussed in the literature, and the reader is referred to the texts by *Battan* [1973], *Doviak & Zrnić* [1984], and *Skolnik* [1990] for more detail.

1.2.4 Wind Profilers

Unlike conventional radars, which detect returns from objects *in* the air, wind-profiling radars make use of signals returned from the air itself. The technique is an outgrowth of research work performed with radars designed to probe the ionosphere; in fact, the first clear-stratosphere vertical wind profiles were made by ionospheric physicists with an ionospheric radar [*Woodman & Guillén*, 1974]. Today’s wind profilers operate in the UHF and VHF bands, observing echoes due to turbulent variations in atmospheric refractive index into the mid-stratosphere and above. Wind velocity can be determined from the Doppler shift of the echoes.

The majority of Chapter 2 is devoted to a review of the theory and implementation of wind profiling techniques.

1.2.5 Other remote sensing systems

These systems are not utilised in the work presented here, but a brief description of each is included for completeness. More detailed information can be found in *Golden et al.* [1986] and *Väisälä* [1988]. In addition, *Browning & Szejwach* [1994] review current and planned remote-sensing systems, with particular emphasis on their application to weather forecasting.

1.2.5.1 Acoustic sounders

Acoustic radars, or sodars, have good vertical resolution but are limited in maximum range by the strong scattering and absorption of sound in the atmosphere. They are used almost exclusively for observations of the boundary layer structure, together with winds in that region. Another application is in the tracking of temperature inversion layers. A major problem with these systems is the severe degradation in performance during rain or high wind conditions—often the times when data are of most interest.

1.2.5.2 RASS: The Radio Acoustic Sounding System

The RASS technique combines an acoustic wave transmitter with a standard electromagnetic wave transmitter (radar). The longitudinal acoustic wave modulates temperature and density, and hence refractive index, as it propagates through the atmosphere. Matching the acoustic wavelength to the radar Bragg wavelength (half the radar wavelength for backscatter) produces enhanced radar backscatter. Since the acoustic wavelength and velocity are temperature dependent, a temperature profile can be constructed by varying the transmitted acoustic frequency and observing the altitude of the enhanced echo. Complications occur due to the strong attenuation of sound in the atmosphere, and the bending of the acoustic wave path by the wind, but careful experiments have proven the utility of the technique [*Matuura et al.*, 1986; *May et al.*, 1989a].

1.2.5.3 Thermodynamic profilers

These are passive radiometric sensors, similar to satellite-borne devices but upward-rather than downward-looking. Radiometers operating in the microwave absorption region for water vapour measure vertically-integrated water content; those operating in the oxygen absorption band measure temperature. Statistical inversion techniques are then used to produce profiles of temperature and humidity [*Hogg et al.*, 1983].

1.2.5.4 Lidar

Lidar is essentially an optical radar, making use of scatter from aerosols and molecules in the atmosphere. There are three main techniques: incoherent lidar, which can measure boundary layer depth and structure; differential absorption lidar, which measures profiles of water vapour, temperature, and gaseous atmospheric constituents; and Doppler lidar, which measures the wind vector in a similar manner to weather radars and wind profilers [*Bilbro et al.*, 1984].

1.3 Motivation and Scope

The standard meteorological techniques of satellite imaging and balloon-borne radiosonde profiles of thermodynamics and wind (which are used as the input into forecasting and general circulation models) are suitable for synoptic scale forecasting, but do not provide data of sufficient detail to thoroughly examine mesoscale phenomena. These smaller-scale events are of significant interest because of their potential for influencing local weather conditions, and their feedback interaction with the mean flow. As the interest in mesoscale phenomena has increased in recent years, so has the search for observing techniques with which to supplement the standard dataset.

Radar wind profilers provide a unique capability to measure vertical wind profiles through the troposphere and into the stratosphere at a time resolution of minutes, and a vertical resolution better than the radiosonde “standard levels” which are usually reported. The measurement is a truer vertical profile than that provided by balloons (which obviously travel with the wind), which is important in the smaller timeframe

of the mesoscale. A further factor which deserves consideration in these times of economic rationalisation is that networks of wind profilers are considerably cheaper to operate than an equivalent network of radiosonde launching sites.

Our primary motivation, then, is to investigate the applicability of radar wind profilers to the observation of mesoscale phenomena. The reasons are twofold: first and foremost, to give added insights to the physics associated with these events; and secondly, to assess the benefits, limitations, and unique capabilities of profilers as applied to the field of mesoscale meteorology.

We begin with a review of wind profiler techniques, and a technical description of the Buckland Park VHF radar. The theory of VHF radar returns from the lower atmosphere, especially scattering mechanisms, is covered in some detail. Balloon-borne radiosondes are used extensively in our observations, and technical matters related to the complete balloon/radiosonde system are examined. The capabilities of the automatic weather station which was on occasion situated at the field site are noted. On a related note, we make a critical assessment of the use of balloon ascent rates as a basis for atmospheric vertical velocity calculations, and question the method's reliability outside of a structured, large-amplitude atmospheric wave environment.

Although studies of the accuracy of radar wind profilers are numerous in the literature, there is a dearth of quantitative information on the subject of profiler altitude coverage and the variations thereof. We extend and generalise the comparisons between radar performance (based on signal-to-noise ratio) and atmospheric characteristics (using thermodynamic profiles) to suggest and test a method of predicting altitude coverage at any site for which nearby historical thermodynamic profiles are available.

Wind profilers are not limited to simply measuring profiles of wind, and we examine some other applications such as stability indices, tropopause height determination, and the possibilities of deriving temperature and humidity profiles from the returned signal. Previous ideas are reviewed, and some improvements are suggested. In particular, a cluster-mean approach to the question of tropopause height retrieval is analysed.

We then move on to a series of observations of mesoscale systems, made during the South Australian Fronts Experiment (SAFEx). The combination of complementary

techniques such as wind profiler and radiosonde observations, surface measurements, satellite imaging, and data assimilation schemes using regional input, allows for an intensive investigation of such systems. In particular, we analyse a severe rain event, identified as a cold-core cyclone with an occluded warm front, and make a critical assessment of the “prefrontal trough” concept as used in Australian-region forecasting. We are also interested in the generation of gravity (buoyancy) waves by strong mesoscale features such as cold fronts, and present some observations of the velocity-perturbation field around these events.

Chapter 2

WIND PROFILER TECHNIQUES

The average Ph.D. thesis is nothing but a transference of bones from one graveyard to another.

— *J. Frank Dobie*

The majority of this thesis is concerned with the analysis and interpretation of data collected by two systems: a VHF radar, and a radiosonde package utilising the Omega navaid network for wind measurement. In this chapter, we review profiler data collection and analysis in some detail, beginning with an overview of VHF radar observations of the atmosphere (Section 2.1) and a discussion of fundamental radar and signal processing techniques (Sections 2.2 and 2.3). Atmospheric scattering mechanisms and profiler wind measurement methods are reviewed in Sections 2.4 and 2.5 respectively.

2.1 Background

Radar (“radio detection and ranging”) techniques rely on the return of a transmitted signal, and hence on the nature of the reflection or scattering target. The earliest radars, operating at frequencies between 1.6 and 20 MHz (in the MF and HF bands, with wavelengths from 200 to 15 metres or so) were able to detect the returns from

enhanced ionisation near 100 km altitude. UHF/SHF weather radars (operating frequencies of a few hundred MHz to a few GHz, and centimetre wavelengths) are sensitive to echoes from hydrometeors (precipitation echoes). Additionally, the use of radar to detect hard targets such as aircraft (or, as the less fortunate reader may have experienced, automobiles) is a well-known application which is common to all systems.

However, it is the study of the clear atmosphere which is of concern here: that is, scattering and reflection from the atmosphere itself, rather than from distinct targets such as free electrons or hydrometeors within the atmosphere. The first reported observations of echoes from the clear air were made with ionospheric radars [*Colwell & Friend*, 1936; *Watson Watt et al.*, 1936], and were associated with large refractive index gradients due to temperature inversions by *Friend* [1949]. With the introduction of GHz radars in the 1940s, and the associated observation of different types of returns from invisible targets (called “angel” echoes), there ensued decades of debate over the nature of clear air returns [*Friend*, 1949; *Plank*, 1954; *Atlas*, 1964; *Atlas*, 1965; *Hardy et al.*, 1966].

By the mid- to late 1960s, the major mechanisms for clear-air returns had been identified as insects and birds at wavelengths less than about 3 cm. and backscatter from strong refractive index fluctuations at longer wavelengths [*Tatarskii*, 1961; *Battan*, 1973]. This led to studies of clear-air turbulence (CAT), including tropopause detection [*Atlas et al.*, 1966b], and the development and breaking of buoyancy waves [*Hicks & Angell*, 1968]. The first Doppler measurements from clear air refractive index fluctuations were reported by *Dobson* [1970]. Another major application of the centimetre scale radars was the observation of clear air convection (see, for example, *Konrad* [1970] and *Harrold & Browning* [1971]).

Interest in longer wavelength (VHF band) radars was sparked by the work at the Jicamarca Radio Observatory in Peru, which resulted in the observation of Doppler radar wind velocities in both the stratosphere and mesosphere [*Woodman & Guillén*, 1974]. The concept of radars capable of observing the atmosphere from the boundary layer up to 100 km or so (called mesosphere-stratosphere-troposphere, or MST radars) produced a flood of projects including: the Sunset facility in Colorado [*Green et al.*, 1975], which was restricted to ST measurements; the SOUSY radar in Germany

[Czechowsky *et al.*, 1976; Rüster & Woodman, 1976; Czechowsky *et al.*, 1984]; and the Poker Flat array in Alaska [Balsley *et al.*, 1979; Balsley *et al.*, 1980]. Röttger & Larsen [1990] (in their Table 5.4) list over twenty individual (M)ST radars (VHF and UHF) across the globe, not including the dedicated components of radar networks. Browning & Szejwach [1994], quoting an unpublished UK Meteorological Office report, list over 170 wind profiling radars of all types in operation worldwide by the end of 1991.

Hardy & Gage [1990] present a comprehensive history of clear-air radar studies, to which the reader is referred for more detail. Before moving on to the more technical sections of this thesis, however, a note on terminology is appropriate. The conventional microwave-band (millimetre–centimetre wavelength) precipitation-sensitive radars will be referred to as *weather radars*. UHF (operating frequencies around 400 MHz, with wavelengths around 75 cm) and VHF (around 50 MHz, with 6 m wavelengths) radars are often labelled “clear-air” radars, which can be misleading; although studies of the clear air through scattering from refractive index fluctuations are their forté, they are also sensitive to precipitation (in general terms, the relative contribution from the hydrometeor component decreases as wavelength increases). We shall refer to meteorological UHF and VHF radars as *wind profiling radars* (or just *wind profilers*), since the measurement of vertical wind profiles is their main application. The term “Doppler radar” is avoided on two counts: firstly, Doppler capabilities are assumed in modern systems, including many weather radars; and secondly, because we are interested in wind-measuring techniques distinct from the common Doppler beam-swinging applications, such as the spaced antenna method (see Section 2.5.2). Also, with meteorological interests confined to the lowest thirty kilometres or so of the atmosphere, a mesospheric capability is not required, and the use of the “MST” label would be misleading for the great majority of meteorological profilers.

2.2 Fundamental Radar Concepts

The techniques used in wind profiler operation are similar to those used with conventional weather radars (as described by Battan [1973], Doviak & Zrnić [1984], and Skolnik [1990]), although profilers employ vertical and near-vertical beams rather than

a low elevation-angle azimuthal scan. Additionally, since clear-air returns are generally weaker than returns from hydrometeors, and have much longer echo coherence times (large fractions of a second as opposed to milliseconds), profilers employ time-domain averaging over successive pulses. This is the basis of the so-called “MST radar technique”, which is applicable to pulsed, coherent radars, and has been described in detail by a number of authors (*e.g.* Balsley [1981], Gossard & Strauch [1983], Röttger [1984], and Röttger & Larsen [1990]). However, there are a number of concepts which are of importance in profiler performance (and hence design), and they will be covered here for future reference. The initial work performed for this thesis involved the implementation of radar controller software and Doppler analysis routines, the scope of which are described in this section and the next.

2.2.1 Pulsed Radar Theory

A radar pulse of duration Δt (the *pulse width*) transmitted from the ground at some time t_1 travels with the speed of light c , and reaches a range r_a after a time $t_a = r_a/c$ (see Figure 2.1). If the pulse is scattered or reflected at that level, energy can return to the transmitter location and be received after a time $t'_1 = 2t_a = 2r_a/c$.

Hence, for monostatic radars (where the transmitting and receiving systems are co-located), the range r to a target can be computed from the round trip time t through:

$$r = ct/2 \quad (2.1)$$

Since the pulse is of finite duration, it is obvious from Figure 2.1 that the volume illuminated at range r_a extends along a range:

$$\Delta r = c\Delta t/2 \quad (2.2)$$

since echoes from the range between $r_a - \Delta r/2$ and r_a reach the receiver simultaneously at time t'_1 . We call Δr the *range resolution*; the smaller Δr , the better we can resolve detail along a profile. The sampled region between $r_a \pm \Delta r/2$ is called a *range gate*.

Radar pulses are transmitted regularly, separated by an interpulse period (IPP) of t_p corresponding to a pulse repetition frequency (PRF) $f_p = 1/t_p$. This allows

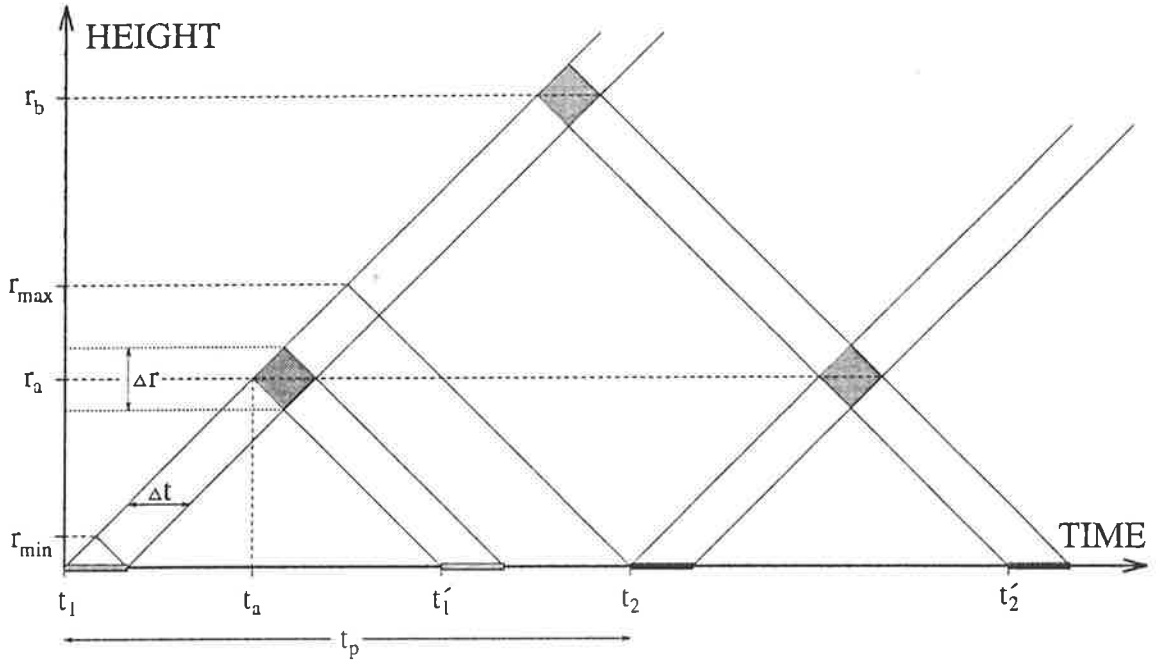


Figure 2.1: A range-time diagram of radar pulse behaviour. The pulse travels with a constant speed c , as do pulse echoes from reflection and scattering layers in the atmosphere. The pulse width is Δt , and pulses are separated by the interpulse period $t_p \gg \Delta t$ (after Röttger [1984]).

us to define the *duty cycle*, $d = \Delta t/t_p = f_p \Delta t$, as a measure of transmitter versus receiver activity (for example, if the transmitter produces a peak power $\mathcal{P}$, then the average power is $d\mathcal{P}$). Pulsing also sets a limit on the maximum range that we can unambiguously observe, due to *range aliasing*. This is illustrated in Figure 2.1, where echoes of the first pulse from range r_b , and the second pulse from range r_a , are both received at time t'_2 , and hence both contribute to the echo power from that range. Since the echo power from the lower range will generally dominate, information from the upper range is obscured. The maximum unambiguous range that we can observe is a function of the interpulse period:

$$r_{\max} = ct_p/2 = c/2f_p \quad (2.3)$$

The lowest range that we can properly sample is determined by the pulse width (see Figure 2.1), and is given by $r_{\min} = \Delta r/2$; that is, half the resolution. In practice, however, transmitter/receiver system recovery effects make it advisable to make the first range gate at an interval no lower than the resolution, Δr .

The two important points from the above treatment are:

- the shorter the pulse, the better the range resolution; and
- the lower the PRF, the better the maximum (non-aliased) range.

Unfortunately, the combination of low PRF and short pulses results in a very low duty cycle, and hence a low average power, which reduces the maximum height that will produce useful (as opposed to possible) echoes. Additionally, short pulses contain a wider spread of frequencies (*i.e.* a greater bandwidth) than long pulses. They are more difficult to produce electronically, are prone to rise-time errors (there are fewer cycles in an individual pulse) and, since receiver bandwidths are generally matched to pulse bandwidths, allow more background noise to enter the receiving system. Pulse coding [Röttger, 1984; Sulzer & Woodman, 1985] can alleviate this problem, but was not implemented on the Buckland Park system and will not be discussed further.

2.2.2 The Radar Equation

In designing a radar system, the echo strength is of prime concern. The *radar equation* relates transmitted and received power for a variety of situations. the most common of which are well known [Tatarskii, 1971; Doviak & Zrnić, 1984; Sato, 1989]. We shall outline the form of the radar equation for three cases of interest here. Firstly, consider the case of an isolated hard target at range r with scattering cross-section ς , and a radar operating at some wavelength λ . If we transmit power $\mathcal{P}_t$ from an antenna of effective area A_e with gain $G = 4\pi A_e/\lambda^2$, then the returned power (after transmission, scattering from the target, and reception) is given by:

$$\mathcal{P}_r = \frac{G\mathcal{P}_t}{4\pi r^2} \frac{\alpha A_e \varsigma}{4\pi r^2} \quad (2.4)$$

$$= \frac{\alpha \mathcal{P}_t A_e^2}{4\pi \lambda^2 r^4} \varsigma \quad (2.5)$$

where α is an efficiency (loss) factor for the system as a whole. Note that the power returned from a hard target varies with an r^{-4} dependence.

Next, consider a homogeneous ensemble of randomly distributed point targets—that is, the case of isotropic volume scatter. Noting that the region illuminated by the radar beam increases with the square of the range, it is not surprising that the radar

equation takes the form [Balsley & Gage, 1980]:

$$\mathcal{P}_r = \frac{\alpha^2 \mathcal{P}_t A_e \Delta r}{8\pi r^2} \eta \quad (2.6)$$

where η is the volume reflectivity; note the r^{-2} dependence.

In a similar manner, the power returned from a single discontinuity by partial (Fresnel) reflection can be determined as [Gage & Balsley, 1980]:

$$\mathcal{P}_r = \frac{\alpha^2 \mathcal{P}_t A_e^2}{4\lambda^2 r^2} |\varrho|^2 \quad (2.7)$$

where ϱ is the amplitude reflection coefficient of the discontinuity. Again, the returned power is proportional to r^{-2} .

Hence, with the appropriate expressions for η and ϱ , Equations 2.6 and 2.7 allow us to specify the magnitude of the returned power. The relationship of η and ϱ to atmospheric structure is described in Section 2.4, for a variety of echoing mechanisms. We finish here by observing that in the real atmosphere, where extreme cases such as the three outlined above are likely to be the exception rather than the rule, the range dependence of power returns is likely to lie somewhere between r^{-2} and r^{-4} . For example, weather radar ground clutter (using narrow beams and low elevation angles) decays as r^{-3} [Doviak & Zrnić, 1984]. The degree to which scatterers fill up the scattering region, as examined by Rastogi & Scheucher [1990], is a significant factor, and will be discussed in Chapter 4.

2.2.3 Doppler Measurements

If there is relative motion between the transmitter and the atmospheric scatterers, the Doppler effect results in a change in the frequency of the returned signal. This frequency change is called the *Doppler shift*. If the target is moving towards the wave source, the observed frequency will be higher (a positive Doppler shift); if it is receding, then the returned frequency is lower (a negative Doppler shift). Noting that a double Doppler shift is encountered in this case—one as the pulse impinges on the scatterers, and another upon scattering—the magnitude of the shift is:

$$f_D = -2V_r/\lambda \quad (2.8)$$

where $V_r = dr/dt$ is the radial velocity, defined as positive in the usual sense (as range increases). For future reference, also note that we can express the Doppler shift as a function of the rate of change of phase ϕ of the returned signal:

$$f_D = \frac{1}{2\pi} \frac{d\phi}{dt} \quad (2.9)$$

Nyquist arguments result in a maximum resolvable Doppler frequency $f_{D_{\max}}$ of half the PRF (*i.e.* $f_p/2$). This corresponds to a maximum measurable (unaliased) radial velocity of:

$$|V_r|_{\max} = \frac{\lambda f_p}{4} = \frac{\lambda c}{8r_{\max}} \quad (2.10)$$

(using Equations 2.3 and 2.8).

Equation 2.10 illustrates a third important point for profiler design: the higher the PRF, the higher the maximum radial velocity that we can observe. In practice it is the effective PRF, or sampling rate, that is important for velocity aliasing, and for that we need to discuss the idea of echo coherence.

2.2.4 Coherency

Typical wind velocities in the troposphere and stratosphere are of the order of a few tens of metres per second in the horizontal, and one or two orders of magnitude smaller for vertical winds. Radial velocities, from standard beam tilts of around ten degrees, are then about 10 ms^{-1} . This corresponds to Doppler shifts of less than 10 Hz with VHF systems (from Equation 2.8). Hence, provided we use a large enough PRF, the radar echo will be heavily oversampled—it's amplitude and phase will vary little from pulse to pulse. This *coherency* in the echoes allows us to improve signal quality by simply adding together a number n_c of sequential samples. This *coherent averaging* reduces the total noise power by removing high-frequency components from the signal, as well as reducing computational load by reducing the total number of sampled points. The *sampling rate* is defined by f_p/n_c , and it determines the time resolution of echo measurements. The largest radial velocity we can measure is given by substitution into Equation 2.10:

$$|V_r|_{\max} = \frac{\lambda f_p}{4n_c} \quad (2.11)$$

The degree of coherent averaging is an important factor in radar experiments, and its selection is ultimately determined by the nature of the turbulent irregularities that are being observed. There is a trade-off between the twin requirements of observing rapidly fading echoes, and obtaining sufficient returned power in an individual sample.

2.2.5 Summary

The final parameter which should be mentioned in the context of profiler performance is the number of points, n_s , which we use to construct a representative time series of echoes in every range gate. Statistical methods are used to analyse radar returns, and these methods obviously benefit from a greater number of points. This must be balanced against physical concerns, primarily the changing state of the atmosphere. More points implies a longer time between samples of the atmosphere, and care must be taken to ensure that the samples are representative. Generally, we would not want more than a couple of minutes of data contributing to any given sample, given a typical Brunt-Väisälä frequency of about five minutes in the troposphere (using Equation 1.9).

In summary, then, the important variables in profiler operation are: the pulse width, which determines the range resolution and affects transmitted power; the PRF, which also affects average transmitted power and limits our maximum range; the number of coherent averages, which restricts our maximum observable velocity and reduces noise power; and the data series length, which limits the time resolution and representativeness of our observations. In combination, they are the major factors which determine the operational performance of a given system outside of hardware concerns. In the next section, we shall discuss the techniques which are used to process the radar returns—the sampled time series of electromagnetic field variations due to scattering from atmospheric refractive index fluctuations.

2.3 Data Analysis Techniques

Analysis of the raw, sampled time series of electromagnetic field variations at the receiving antenna on the ground is the first step in determining meteorologically important parameters. In practice, this analysis can be carried out either in the frequency or

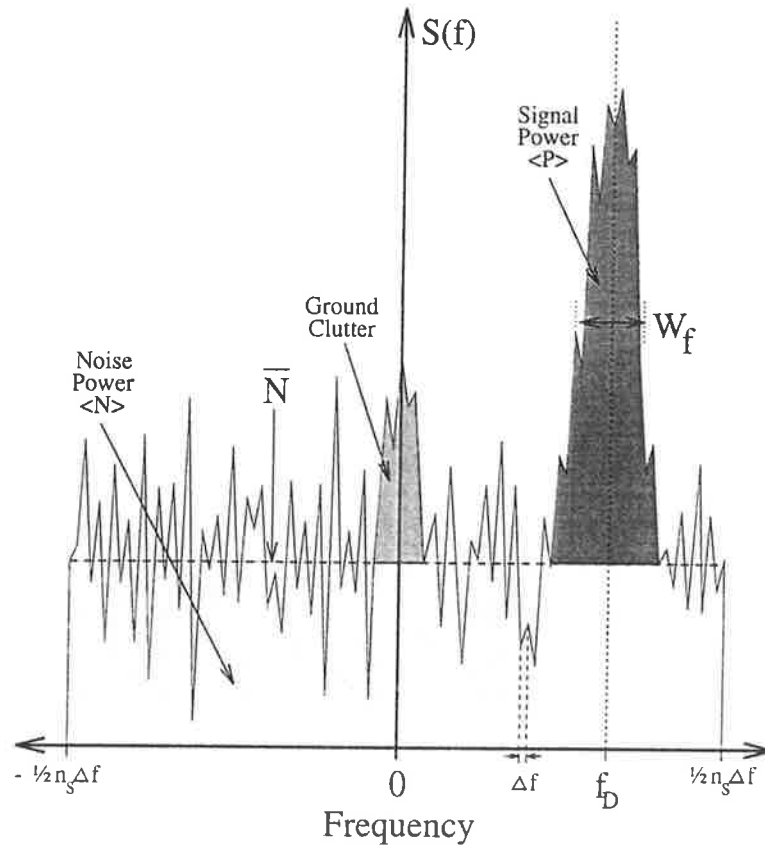


Figure 2.2: A typical power spectrum showing measurement parameters.

time domain, each of which has advantages and disadvantages which will be discussed next.

2.3.1 Frequency Domain Analysis

The reduction of the pulse-to-pulse time-varying echo strength time series to its physically important characteristics is perhaps most readily appreciated in the frequency domain. Note that if a single discrete waveform is Fourier transformed, the output frequency spectrum will show a Doppler shift with ambiguous sign. To resolve this, the returns from each transmitted pulse are mixed with sine and cosine phase-locked signals from a local oscillator. The resultant signals are called the *in-phase* and *quadrature* components respectively, and together allow us to construct a complex time series, z_i , which can be unambiguously Fourier decomposed.

A typical power spectrum is shown in Figure 2.2. The central frequency corresponds to the transmitted frequency, with Doppler shifted frequencies on either side. As

described in Section 2.2.3, a positive [negative] Doppler shift corresponds to motion towards [away from] the profiler. Four basic quantities (moments) can be measured from the spectrum: the noise power $\langle N \rangle$, or equivalently the average background signal level $\bar{N}$; the signal power $\langle P \rangle$, given by the area under the peak above the noise level; the mean Doppler shift f_D , which can be converted to a velocity using Equation 2.8; and the width W_f of the peak, which is related to the degree of turbulence in the sampled volume (see, for example, *Hocking* [1983a]). Another feature which can appear is a central peak, caused by reflections from stationary targets, usually in sidelobes of the beam; this is called *ground clutter*. Other contaminations to the spectrum can be caused by aliasing, variable interference, ocean wave echoes, or mixed scattering and reflection modes; their elimination is discussed by, for example, *Röttger* [1980a], *Rastogi & Holt* [1981], *Rastogi & Röttger* [1982], *Hocking* [1983a], and *Hocking* [1983b]. A useful derived parameter is the signal-to-noise ratio (SNR), defined by $\langle P \rangle / \langle N \rangle$. An alternative definition uses only the noise within the signal bandwidth, which essentially scales the SNR by a factor $n_s \Delta f / W_f$.

Mathematical details of spectral moment estimation can be found in the reviews by *Zrnić* [1979], *Doviak & Zrnić* [1984] and *Woodman* [1985]. Generally, however, computation in the frequency domain allows the removal of spectral artifacts, as well as the identification of multimodal spectra (such as combined clear-air and precipitation echoes). If the sampled spectra are free of this type of contamination, then analysis in the time domain is usually more efficient.

2.3.2 Time Domain Analysis

The power spectrum and autocorrelation function of a sampled signal comprise a discrete Fourier transform pair (the discrete Wiener-Khinchine theorem). Hence, they contain identical information about the lower spectral moments, distributed over frequency in the power spectrum, and concentrated in the small lags of the autocorrelation function [*Woodman & Guillén*, 1974; *Bowhill*, 1983]. We define the m -th lag value of the autocorrelation function of the N -point complex time series z_i by:

$$R(m) = \frac{1}{N-m} \sum_{k=0}^{N-m-1} z_k^* z_{k+m} \quad (2.12)$$

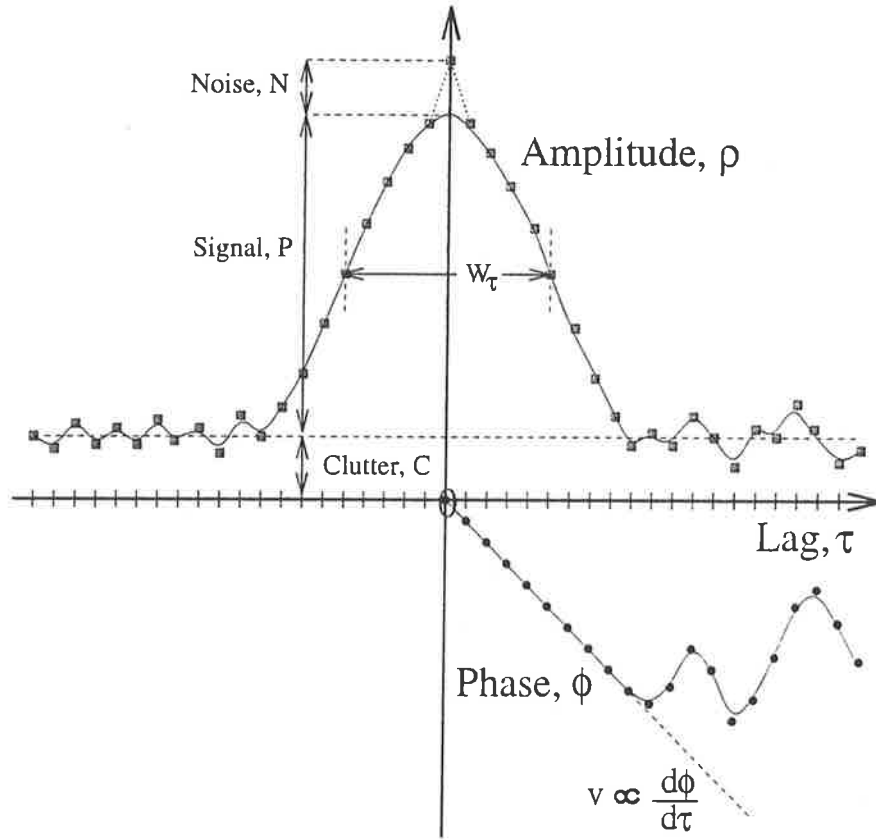


Figure 2.3: A typical autocorrelation function showing measurement parameters.

where $*$ denotes conjugation. In fact, there is a direct relationship between the n -th classical spectral moment M_n and the n -th derivative of the autocorrelation function at zero lag:

$$M_n = R^{[n]}(0)/(2\pi j)^n \quad (2.13)$$

A typical autocorrelation function is shown in Figure 2.3, displayed as amplitude ρ and phase ϕ components as a function of the lag, τ . Here, random background noise (which decorrelates rapidly) appears as a “spike” of size N in amplitude at zero lag, while ground clutter produces a nearly constant amplitude platform (C) upon which is superimposed the expected (pseudo-) Gaussian. The interpolated zero-lag value of this Gaussian, corrected for clutter, is a direct measure of the signal power P , while its width W_τ is a measure of the echo lifetime (and, under certain conditions, the turbulent velocity variance). As an aside, it is advisable to set the time series length n_s to a value between 25 and 50 times the autocorrelation width, to reduce the effects of small-scale temporal variability whilst retaining a representative data

record [*Hocking*, 1983*b*; *May*, 1988]. Finally, the mean frequency shift (and hence the Doppler velocity from Equation 2.8) is given by the slope of the autocorrelation phase (Equation 2.9). The SNR is defined by P/N , and is usually expressed in decibel (dB) units by taking $10\log_{10}(P/N)$.

The single delay autocorrelation approach estimates signal power and Doppler velocity from the zero- and first-lag values of ρ and ϕ respectively. *Woodman* [1985] gives a detailed evaluation of this method, in particular comparing it to multiple delay techniques, and concludes that provided the signal-to-noise ratio (SNR) is sufficiently high (greater than about 0 dB), the single delay approach is a simple, accurate, and statistically efficient estimator. It is also computationally efficient, a factor which is of less importance today than a few years ago; however, computational limitations were a significant motivation in the design of the original Buckland Park profiler data acquisition software. Radar data from Doppler technique observations presented in this thesis are analysed using this approach, although autocorrelation widths are estimated via a spline-fitted Gaussian.

2.4 Scattering Mechanisms

The nature of atmospheric scattering at radio wavelengths has been discussed at length in the literature over the last ten to fifteen years. The state of knowledge has been detailed in reviews by *Gage & Balsley* [1980], and more recently *Gage* [1990]. Our aim here is to present a brief summary of the field, and introduce the groundwork which is later used extensively in our methods and interpretations.

Radar echoes at UHF and VHF are due to fluctuations in the radio refractive index of the atmosphere, n , expressed as:

$$n - 1 = (0.373)e/T^2 + (77.6 \times 10^{-6})p/T - 40.3N_e/f^2 \quad (2.14)$$

where e , p , and T are the partial pressure of water vapour (hPa), atmospheric pressure (hPa), and temperature (K) respectively. N_e is the number density of free electrons (m^{-3}), and f is the radar operating frequency (MHz) (*e.g.* *Röttger* [1984]). The third (“free electron”) term is of negligible importance below about 50 km altitude; hence,

we apply only the first (“wet”) and second (“dry”) terms in what follows.

Spatial fluctuations in n which match the Bragg condition, possessing scales of half the radar wavelength, will produce scatter and reflection of incident energy. These fluctuations or irregularities are generally produced by turbulence (possibly with some degree of horizontal coherence), small-scale waves, or a combination of both. The most appropriate parameter to specify the intensity of refractivity turbulence is the turbulence refractive index structure constant, C_n^2 , defined over a layer δr by [Ottersten, 1969a]:

$$C_n^2(r_0) = \overline{[n(r_0 + \delta r) - n(r_0)]}(\delta r)^{-2/3} \quad (2.15)$$

It is convenient at this point to introduce the vertical gradient of refractive index, M , given by [Ottersten, 1969b]:

$$M = -77.6 \times 10^{-6} \frac{p}{T} \left(\frac{\partial \ln \theta}{\partial z} \right) \times \left[1 + \frac{15500q}{T} \left(1 - \frac{1}{2} \frac{\frac{\partial \ln q}{\partial z}}{\frac{\partial \ln \theta}{\partial z}} \right) \right] \quad (2.16)$$

where θ is the potential temperature (K) and q is the specific humidity. Note that we could express Equation 2.16 as:

$$M \equiv M_d \times [1 + \dots] \quad (2.17)$$

In the stratosphere and upper troposphere, when q is small enough, the bracket in Equation 2.17 can be set to unity, and we could deal with just the “dry” part, M_d :

$$M_d = -77.6 \times 10^{-6} \frac{p}{T} \left(\frac{\partial \ln \theta}{\partial z} \right) \quad (2.18)$$

Referring to Equation 1.8, it is immediately apparent that M and M_d are proportional to N^2 , the Brunt-Väisälä (buoyancy) frequency, and are thus measures of static stability in the atmosphere. In an atmosphere where the environmental lapse rate equals the dry adiabatic lapse rate, $\partial \ln \theta / \partial z = 0$ (from Equations 1.8 and 1.9), and hence M vanishes. We shall examine the implications of this result later in this section.

Both C_n^2 and M^2 are important in the modelling of radar returns from the atmosphere via the radar equation (Equations 2.6 and 2.7). The radar reflectivity for scattering from volume-filling isotropic turbulence (given Bragg matching) is [Ottersten, 1969b]:

$$\eta = 0.38 C_n^2 \lambda^{-1/3} \quad (2.19)$$

The theory of *Tatarskii* [1971] relates C_n^2 to M^2 through:

$$C_n^2 = a^2 \alpha' L_o^{4/3} M^2 \quad (2.20)$$

where a^2 is a universal constant close to 2.8, α' is a ratio of eddy diffusivities usually taken to be unity, and L_o is the outer scale of turbulence, around 10–100 m in the lower-middle atmosphere [*Warnock & VanZandt*, 1985].

The *Tatarskii* approach assumes low transverse coherence across the radar beam, with a correlation distance l_t less than the Fresnel zone radius $r_F = \sqrt{r\lambda/2}$. Such isotropic turbulence is considered to be the major mechanism for radar backscatter at UHF, and at VHF for large enough zenith angles χ (greater than 10° or so). Since backscattered power is proportional to η , which is proportional to C_n^2 , the measurement of returned power allows the estimation of profiles of C_n^2 , although a properly calibrated radar system is required if absolute C_n^2 values are to be reported [*Nastrom et al.*, 1986; *Eaton et al.*, 1988; *Frisch & Weber*, 1992]. A number of radar experiments have established the relevance of turbulent scatter in these cases (*e.g.* *Woodman & Guillén* [1974], *VanZandt et al.* [1978], *Ecklund et al.* [1979]). A series of papers [*VanZandt et al.*, 1978; *VanZandt et al.*, 1981; *Warnock & VanZandt*, 1985] outlines improvements in the modelling of C_n^2 profiles from archival rawinsonde measurements of M^2 , although application is limited to backscatter from hydrostatically stable regions of the free atmosphere where turbulence arises from shear instability.

Isotropic turbulence does not explain the entirety of radar observations, especially at VHF. For example, experiments [*Röttger et al.*, 1981; *Tsuda et al.*, 1986] have shown that returned signal power decreases with zenith angle, indicating a degree of horizontal anisotropy (or stratification, or *aspect sensitivity*) in the scattering medium. The decrease is estimated at approximately 1–2 dB per degree in the troposphere, and 2–3 dB per degree in the stratosphere. The concept of scatter from anisotropic turbulence [*Staras*, 1955; *Gage & Balsley*, 1980], where correlation distances vary in three dimensions, has been shown to be consistent with observations at some times [*Waterman et al.*, 1985], although in general it predicts a rate of power decrease with angle which slower than generally observed [*Tsuda et al.*, 1986]. Alternatively, partial (Fresnel) reflection processes can be modelled (Equation 2.7), assuming that transverse

coherency extends over a Fresnel zone [Röttger & Liu, 1978]; that is, proposing the existence of a coherent layer of concentrated refractive index gradient. Calculation of the reflection coefficient, $|\rho^2|$, in this case is complicated by the need for fine-scale knowledge of the refractive index profile, although the returns will obviously maximise at normal incidence. The specular nature of partial-reflection returns will cause the echoes to have a longer coherence time (a narrower Doppler spectrum) than echoes due to turbulence (see, for example, Sheen *et al.* [1985]).

Perhaps a more realistic model of the specular nature of radar returns from (near-) vertical incidence is Fresnel scatter [Gage *et al.*, 1981], which postulates the existence of several horizontally-coherent, vertically-incoherent refractive index discontinuities within a given range gate and across distances the order of a Fresnel zone. Fresnel scatter echoes will have a distinct aspect sensitivity, but have temporal characteristics similar to turbulent scatter since the discontinuities are statistically independent. Observations of backscattered power proportional to radar pulse width Δr [Hocking & Röttger, 1983; Gage *et al.*, 1985] are consistent with such multiple layer partial reflection, for which the amplitude reflection coefficient is [Gage, 1990]:

$$|\rho^2| = F(2k, r)^2 \Delta r \overline{M^2} / 4 \quad (2.21)$$

where:

$$F(2k, r)^2 = 4k^2 E_\zeta(2k, r) \quad (2.22)$$

Here, $E_\zeta(k)$ is the spectrum of Bragg-scale vertical displacements in the atmosphere, related to $\Phi_n(k)$, the vertical wavenumber (k) spectrum of generalised potential refractive index, by $\Phi_n(k) = E_\zeta \overline{M^2}$. Studies comparing power returns on vertical VHF beams with this model (*e.g.* Tsuda *et al.* [1988]) have demonstrated excellent agreement, and shown that the vertical profile of backscattered power is dominated by the vertical profile of $\overline{M^2}$. The altitude (r) dependence of F is uncertain: the observations of Balsley & Gage [1981] could be interpreted as an r^{-2} proportionality; Gage *et al.* [1985] propose an exponential dependence; while Tsuda *et al.* [1988] find a uniform distribution with height. The work of Hocking & Röttger [1983] also implies that some variation from Δr -proportionality is possible if only a few layers within the scattering volume contribute to the returned power, or if M varies widely with altitude.

Of course, it is in combination that isotropic and anisotropic turbulence, Fresnel reflection, and Fresnel scatter contribute to the returned power. Measurements of aspect sensitivity, fading and coherence statistics, and range dependence allow the identification of particular mechanisms contributing to a given set of observations. Although a common feature appears to be a proportionality to M^2 , the physical interpretation differs: the scale of the predominant fluctuations in M^2 determines the nature of the scattering process. Further to this, since M vanishes under adiabatic lapse rates, we note that returned power will generally tend towards zero as the atmosphere becomes adiabatic. Naturally, variations of M around zero result in an M^2 that will be small and non-zero, so a fraction of the transmitted power will almost always be returned—whether it can be detected is another matter. The small returns can be increased by a number of methods which revolve around increasing the effective transmitted power. Adiabatic lapse rates occur most frequently in the upper troposphere and, together with the characteristic dryness of that region, will tend to produce low power returns. This is examined experimentally in Chapter 4.

2.5 The Measurement of Winds

The measurement of the three-dimensional wind field is the major meteorological application of VHF/UHF radars. There are two main methods in operational use, both of which assume the Taylor hypothesis of scattering irregularities moving with the mean wind. The first, and most widely applied, is Doppler beam swinging (DBS), where the echoed Doppler-shifted frequency is measured with a narrow beam pointed in different directions. The other method involves cross-correlating the received signals at three or more spaced antennas to determine the motion of the scatterers via the movement of the diffraction pattern on the ground. Other techniques, such as Doppler interferometry [Farley *et al.*, 1981; Röttger, 1984; Adams *et al.*, 1985; Röttger & Ierkic, 1985; Adams *et al.*, 1986], have been applied on occasion to the measurement of winds, but will not be reviewed herein in favour of the major—and applied—techniques.

2.5.1 Doppler Techniques

As discussed above, autocorrelation or spectral analysis can be used to measure the radial velocity of scatterers (*May et al.* [1989b] and *May & Strauch* [1989] discuss the accuracy of various estimators for the Doppler shift in the estimation of winds). If we assume the Taylor hypothesis, that the scatterers move with the three-dimensional mean wind $\mathbf{u} = (u, v, w)$, then a beam pointed to an azimuth angle ψ (clockwise from north) and zenith angle χ and will measure a radial velocity:

$$V_r(\psi, \chi) = u \sin \psi \sin \chi + v \cos \psi \sin \chi + w \cos \chi \quad (2.23)$$

If the wind field is uniform in space, and the zenith angle is held constant, V_r will vary sinusoidally with azimuth. Fourier analysis of $V_r(\psi)$ allows determination of $\mathbf{u}$; this is the velocity azimuth display (VAD) technique.

In most radar applications, the VAD method is simplified by the use of three beam directions: one vertical beam ($\chi = 0$), and two off-vertical but orthogonal beams, usually directed north and east. This reduces Equation 2.23 to three equations in the three wind components, which can be solved simultaneously to determine $\mathbf{u}$. This is known as the three-beam method, and is one of a number of fixed-beam techniques. If we are willing to assume that the vertical velocity, w , is negligible, it can be simplified to a two-beam method to determine u and v ; if we wish to take account of spatial variability, we can make use of five or more beams (one vertical, and four or more off-vertical). The latter method also enables us to calculate the divergence of the wind field [*Clark et al.*, 1986]. Alternatively, more accurate estimates of the vertical velocity can be obtained by keeping the azimuth constant, and changing the elevation angle. This is known as the velocity elevation display (VED) technique, and it can be used to measure the full three-dimensional wind field by extension to multiple azimuth angles. VED measurements are also employed to characterise the aspect sensitivity of scatterers, as described in Section 2.4.

Koscielny et al. [1984] compare VAD, VED, and fixed beam techniques, together with the uncertainties and biases involved in using a limited number of beam directions. One problem common to any radar beam system, however, involves the beam width and associated wind shear broadening of the measured spectrum [*Hocking*, 1983a].

Together with the spectral broadening due to turbulent velocity fluctuations within the range gate volume, a finite beam width means that radial velocity components vary across the beam, contributing towards further broadening; wind shear through the beam volume produces the same effect. Additionally, the product of the antenna beam pattern with the angular distribution of aspect sensitivity produces an effective beam direction biased towards the zenith [Röttger, 1980a], which will result in an underestimation of velocity. It is apparent that narrow antenna beams are an advantage in reducing these contaminations.

The other major decision in the design of DBS systems is the choice of zenith angle χ . The arguments in favour of small versus large zenith angles are summarised by Röttger [1984] and Röttger & Larsen [1990]: essentially, the increase in echo strength due to anisotropies at near-vertical incidence, together with spatial uniformity and range-sensitivity considerations, encourage the selection of small χ despite the decreased relative error in velocity measurements and the reduction of pointing-angle errors at larger off-zenith angles. Also, note that increasing χ results in an increased radial velocity measurement, which requires a rapid sampling rate (from Equation 2.11); this is associated with measurement problems, as outlined earlier in this chapter. Generally, zenith angles near ten to fifteen degrees are considered to be an effective compromise [Strauch, 1983; Strauch *et al.*, 1984; Tsuda *et al.*, 1986].

DBS winds have been extensively compared with balloon-borne sondes (both radar-tracked rawinsondes, and navaid systems) at every radar site, ranging from small-scale tests [Gage & Balsley, 1978; Warnock *et al.*, 1978] to more extensive, longer duration experiments involving a large dataset [Fukao *et al.*, 1982; Weber & Wuertz, 1990; May, 1993b]. Frisch *et al.* [1986] make an intercomparison of UHF and VHF Doppler winds, and report an almost perfect correlation, while Gage *et al.* [1988] favourably compare radar winds from a tropical site with meteorological service synoptic analyses. Comparisons with aircraft winds have also been successfully performed [Green *et al.*, 1980; Angevine & MacPherson, 1995]. It is generally found that the radar winds are within the joint error ranges of the techniques, or that differences can be ascribed to spatial differences between the observed regions [Jasperson, 1982]. For example, an RMS wind difference of $2\text{--}4\text{ ms}^{-1}$ is typical for the radar-sonde comparisons referenced

above, which compares well with the errors in simultaneous sonde ascents [Kessler *et al.*, 1985] and DBS estimates [Strauch *et al.*, 1987].

2.5.2 The Spaced Antenna Technique

The history of the spaced antenna (SA) technique has recently been reviewed by Briggs [1993]; the theoretical details are presented by Briggs [1984]. Essentially, the moving diffraction pattern on the ground produced by a vertically directed radar beam scattering off irregularities in the atmosphere is sampled at a number of spaced receiving antennas. Cross-correlation of the signals at each antenna allows measurement of the time delays between antennas, which can be used to calculate the “apparent” velocity of the scatterers. Application of the full correlation analysis (FCA) [Briggs, 1968*a*; Briggs, 1968*b*] corrects for the effects of turbulence (non-stationarity of the scatterers) to produce the “true” velocity. The FCA also yields a number of other useful parameters, mainly related to the variability of the scatterers, which are reviewed by Hocking *et al.* [1989].

Briggs [1980] and May [1990] compare the SA and DBS techniques, and demonstrate that although the mathematics of data processing are quite different, both techniques derive their information from the amplitudes and phases of radar returns from different angles, and are essentially equivalent. However, there are a number of practical advantages to the SA technique:

- DBS techniques require a narrow beamwidth to accurately define the radial direction in which velocity is measured. This generally requires a large, steerable antenna. The SA technique makes use of the variation of Doppler shift with angle, and requires beamwidths of at least a few degrees, directed vertically;
- the SA method measures the wind in a single region above the radar, whereas DBS techniques measure components of the wind in regions which are separated in the horizontal by a considerable distance;
- the use of vertical beams in the SA technique allows us to take advantage of increased power returns due to anisotropic (aspect sensitive) scatterers.

The choice of antenna spacing is important in SA systems, corresponding to the choice of zenith angle in DBS systems. Given typical echo coherency times of about one second, and mean lower-atmosphere winds of about 20 ms^{-1} , an antenna spacing of a few tens of metres on the ground ensures that echoes should remain coherent across the array. Alternatively we can consider the *pattern scale*—the spatial lag for the autocorrelation function to fall to 0.5—which is a measure of spatial coherency and provides information on the degree of specularity of the atmospheric scatterers (small pattern scales are associated with isotropic scatter, larger scales with more specular mechanisms). The antenna spacing should be comparable to the mean pattern scale, which presents some difficulties since pattern scale varies in altitude and time, but antenna spacing is usually fixed. Optimum spacings must be determined experimentally, although studies show that tropospheric and lower stratospheric VHF pattern scales of around 50 m are typical [Vincent *et al.*, 1987; Van Baelen *et al.*, 1990]). The observed pattern scale is also dependent on the beamwidth of the transmitting antenna: essentially, returns are a product of the antenna beam pattern and the aspect sensitivity of the scatterers (see, for example, Hocking [1985]). Diffraction theory tells us that an infinitely narrow beam will produce a near-constant amplitude distribution on the ground, or equivalently a large “pattern scale” independent of the scattering irregularities; illuminating the whole sky (with a very wide beam) would produce returns totally dependent on the nature of the scatterers. The Buckland Park VHF profiler transmitting antenna beamwidth of around 3° is about as narrow as one would envisage using in practice for spaced antenna studies.

Of course, the sampling at each antenna must be rapid enough to resolve the correlation time; the results of neglecting this concern are discussed in Section 3.4), and are addressed in the modelling study of Holdsworth & Reid [1995a]. Errors in wind estimates made with the SA technique are discussed by May [1988] (considering the statistical variation in derived parameters) and Holdsworth & Reid [1995a] (modelling the effects of sampling rate and antenna spacing). It should also be noted that there are a number of rejection criteria employed in the FCA, as described by Briggs [1984]. They include checks on signal strength and fading, phase consistency around the antennas, and overly large apparent-to-true velocity corrections. The use of these

criteria significantly improves the reliability of wind measurements, since many spurious estimates will be rejected without contributing to the wind record, reducing the occurrence of outliers in the dataset. *Vincent et al.* [1987] quote precisions of about 1 ms^{-1} in hourly-average independent FCA-derived wind analyses during quiet periods, demonstrating reproducibility: again, poor quality estimates tend to be rejected.

Spaced antenna wind estimates throughout the atmosphere have been compared with a number of other techniques in an effort to ascertain accuracy. Upper atmosphere comparisons have been made with rocket [*Vincent et al.*, 1977], meteor drift [*Cervera & Reid*, 1995], and DBS [*Reid*, 1988] wind measurements. Comparisons with radiosondes in the lower atmosphere have been made by *Vincent & Röttger* [1980], while *Röttger & Czechowsky* [1980] and *Van Baelen et al.* [1990] have compared SA winds with DBS estimates. *Vincent et al.* [1987] made an intercomparison of all three techniques (SA, DBS, and radiosonde), while *Sheppard et al.* [1993] made a comparison of interferometric winds with SA estimates. In passing, we note the concerns raised by *Holdsworth & Reid* [1995b] and *Vandepeer & Reid* [1995] with regards to interferometric and imaging techniques, and the inherent problems and biases therein which require careful consideration when interpreting comparisons. If the effects of turbulent motion and receiver noise are disregarded, the derived interferometric wind is theoretically equivalent to the spaced antenna *apparent* velocity. All comparisons to date have shown a good agreement, although SA wind magnitudes appear to have a slight ($\sim 5\%$) systematic bias towards underestimation. This is addressed by *Holdsworth & Reid* [1995a] and *Holdsworth & Reid* [1995b] from a modelling perspective, and the majority of estimation errors are found to be associated with uncompensated receiver noise and inappropriate sampling rates. We present our own comparisons in the following chapter (Section 3.4), which are the first extensive comparisons of SA wind measurements with co-located radiosonde launches (minimising the effects of spatial separation between sondes and sampled radar volume).

Chapter 3

OBSERVING PLATFORMS AND INITIAL CONSIDERATIONS

May every young scientist remember ... and not fail to keep his eyes open for the possibility that an irritating failure of his apparatus to give consistent results may once or twice in a lifetime conceal an important discovery.

— *Patrick Blackett*

Having reviewed wind profiler techniques in Chapter 2, we move on to an examination of the data collection platforms and some initial results which are important with regards to future interpretation. The details of the Buckland Park VHF radar are discussed in Section 3.1, including an assessment of errors in the radar phasing subsystems and the impact of such errors on wind measurements. Technical details of the Väisälä MARWINsonde system and automatic weather station which were used to make a large number of the measurements presented in this thesis are presented in Sections 3.2 and 3.3 respectively.

An intercomparison of radar and sonde winds is made in Section 3.4; this represents the first extensive comparison of spaced antenna winds with co-located radiosonde measurements, with the aim of reducing differences due to spatial variability. The concept of an averaging technique which minimises differences due to temporal variability is also introduced. Finally, the measurement of atmospheric vertical velocity

is treated in some detail in Section 3.5, including a critical examination of the use of balloon ascent rates in estimating vertical velocities (Section 3.5.2).

3.1 The Buckland Park VHF Radar

3.1.1 Hardware and Operating Parameters

The 54.1 MHz VHF radar is located at Buckland Park ($34^{\circ}37'S$, $138^{\circ}29'E$), about 35 km north of Adelaide ($34^{\circ}55'S$, $138^{\circ}35'E$) on a flat coastal plain, and is some 50 km upstream from any significant orography. The transmitting array (depicted in Figure 3.1) is approximately 100 m square, and consists of 32 coaxial-collinear (CoCo) dipole rows. There are two sets of sixteen 48-element dipole rows with $\lambda/2$ spacing, separated by λ (essentially a “missing” row where the transmitting and phasing equipment is located), and suspended $\lambda/4$ above the ground [Balsley & Ecklund, 1972]. The ground plane is improved by $\lambda/12$ -spaced copper wires on the ground. Operating in monostatic Doppler mode, it is capable of observing the atmosphere from a lower altitude limit of 2 km (because of recovery time requirements in the transmit/receive switching hardware) up to the mid-stratosphere, every few hundred metres in the vertical. The beam can be directed vertically, or swung 11° off zenith towards the east and west by inserting phasing cables between the transmitter and each dipole row. In addition, three Yagi receiving arrays, each of sixteen 3-element antennas, adjacent to the main array allow spaced antenna measurements of the full three-dimensional wind field. The reader is referred to Vincent *et al.* [1987] for more details; typical operating parameters during the tenure of this thesis are summarised in Table 3.1.

The default operating mode of the profiler usually involved a five minute cycle, alternating one vertical Doppler retrieval (64 seconds of data collection, plus analysis time) with one spaced antenna run (128 seconds plus analysis). This juxtaposition of cycles enabled us to take advantage of the narrow Doppler beam in estimating vertical velocities, together with the enhanced power returns of the SA method for horizontal wind estimations. Measurements could not be made simultaneously, as only three receiving channels were available; switching between reception on the large

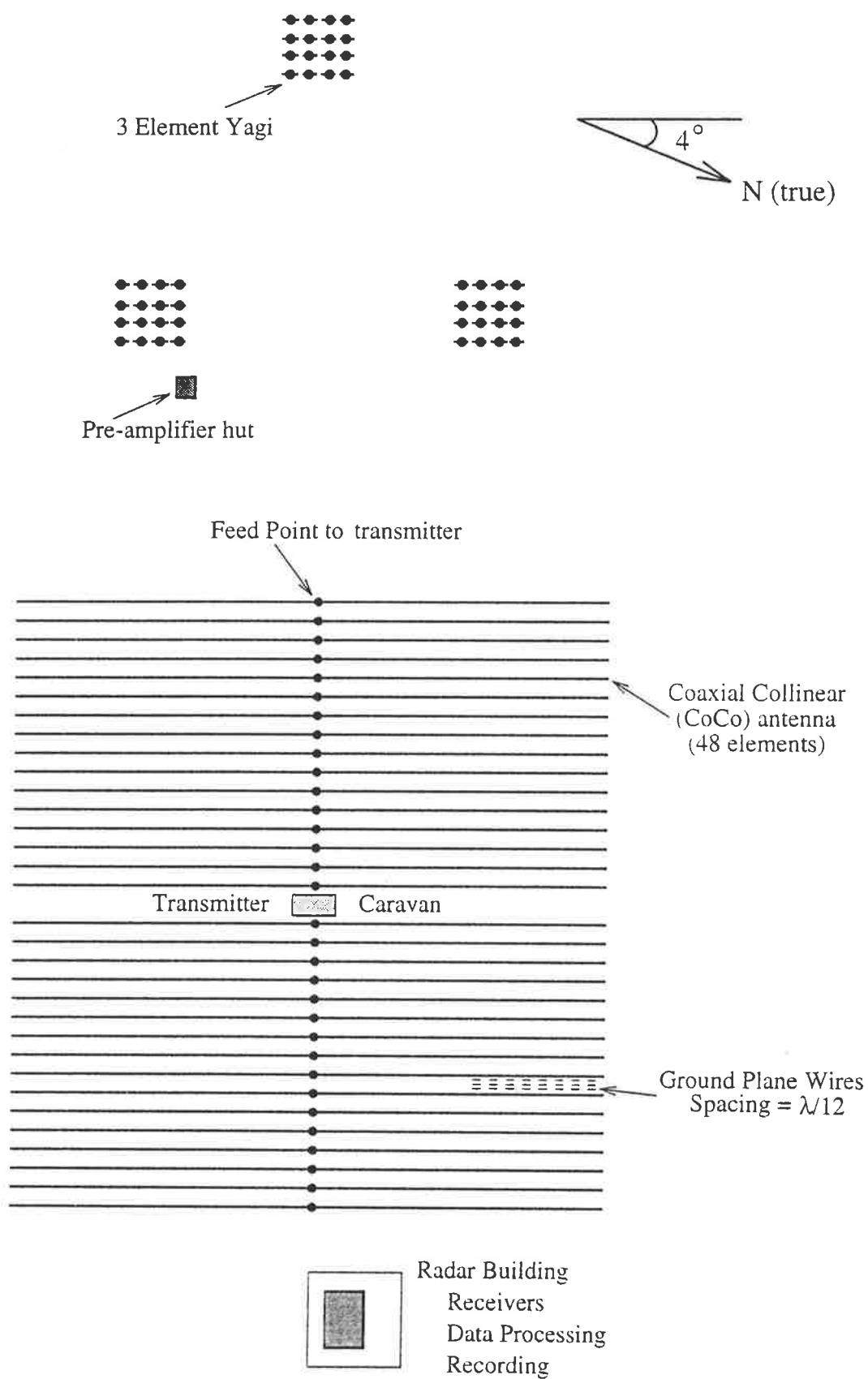


Figure 3.1: A schematic of the Buckland Park VHF radar.

Table 3.1: Operational parameters of the Buckland Park VHF radar in Doppler (DBS) and spaced antenna (SA) modes, appropriate to the period August–October 1992. Present-day peak transmitter power output is 32 kW.

Radar system:	Monostatic DBS/bistatic SA pulse radar
Operational frequency:	54.1 MHz ($\lambda = 5.55$ m)
Transmission antenna (Tx):	Coaxial-collinear dipole array (32 rows of 48 $\lambda/2$ elements)
Tx antenna area:	7900 m ²
Tx antenna efficiency:	≈ 0.3
Tx beamwidth:	1.6° half-power half-width
Beam steerability (ψ, χ):	Vertical (0°, 0°), East (86°, 11°), West (266°, 11°)
Peak Tx power:	16 kW (from eight, 2 kW solid-state Tx modules)
DBS reception (Rx):	On CoCo array, parameters as above
Pulse length:	6.7 μ s
Range resolution:	1000 m, but oversampled every 500 m
Range coverage:	2.0–19.5 km inclusive (36 range gates)
PRF:	4096 Hz
Average Tx power:	440 W (2.7% duty cycle)
Coherent integrations:	1024 points
Effective sample rate:	4 Hz (intersample period 0.25 seconds)
Record length:	64 seconds (256 points)
Maximum unaliased velocity:	22.2 ms ⁻¹
SA reception (Rx):	3 Yagi arrays of sixteen 0.7 λ -spaced 3-element Yagi antennas
SA Rx effective area:	≈ 150 m ² (each Yagi array)
SA Rx spacing:	50 m equilateral triangle, at 176° and 236° from Yagi array Rx0
SA Rx beamwidth:	9° half-power half-width
Pulse length:	6.7 μ s
Range resolution:	1000 m, but oversampled every 500 m
Range coverage:	2.0–14.5 km inclusive (26 range gates)
PRF:	8192 Hz
Average Tx power:	880 W (5.4% duty cycle)
Coherent integrations:	1024 points in hardware, then 4 points in software
Effective sample rate:	2 Hz (intersample period 0.5 seconds)
Record length:	128 seconds (256 points)

CoCo array (for Doppler measurements) and Yagi array Rx0 was performed via a simple logic chip setup, parasitic to the beam direction logic code [Low, 1990]. The details are given in Appendix A.

3.1.2 Antenna Array Phasing

After the rewriting, and subsequent implementation and testing, of the radar controller software, an extensive mechanical and electrical refurbishment of the Buckland Park array was undertaken (collaboratively with M. A. Cervera). Two areas of this work bear mention, and are discussed next.

3.1.2.1 Beam Pointing

As mentioned earlier, the beam pointing angle is an important parameter to know. The Buckland Park site is level to better than 0.1° , but phase delays during propagation through cables—especially an error in power transfer cable lengths—could conceivably produce a bias. Hence the question, how close to vertical is the “vertical” beam, and how close to the desired 11° are the cable-delayed off-vertical beams? A phase-locked signal was fed sequentially into each coaxial-collinear row from the first common point after the transmitter modules, and the resultant signal (after propagation through power splitters, the phasing system, and feeder cables) was measured at the centre feed point on each row. The results are displayed in Figure 3.2 for three cases: a vertical beam setup, with phasing cables bypassed; and two off-vertical beams, with the 11° cables included. Note that simple geometry implies that the phase delay δ required between adjacent rows separated by $l = \lambda/2$, to produce a beam at zenith angle χ is given by:

$$\delta = \frac{2\pi}{\lambda} l \sin \chi = \pi \sin \chi \quad (3.1)$$

The 11° beam is formed in practice by the addition of two sets of cables (cut for 4° and 7°), using $\sin \chi \approx \chi$ (small χ), which is accurate to better than 0.05° here; however, the modelling results referred to below do use the actual cable lengths.

Figure 3.2 shows the measured phases, linear least-squares fits to the measurements, and indicates the predicted model slopes from Equation 3.1: the numerical results are

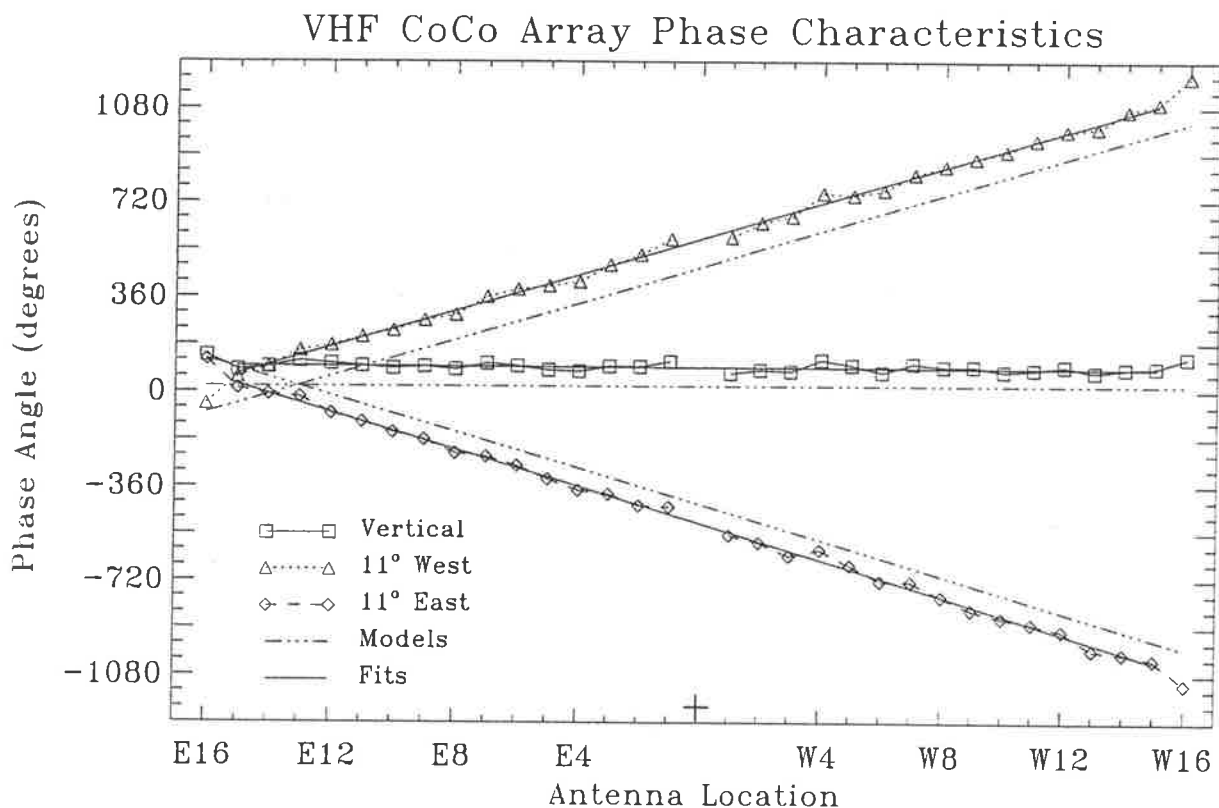


Figure 3.2: VHF CoCo antenna phase slopes across the array for vertical and 11° East and West off-vertical beams. The position of the missing antenna at the location of the transmitting caravan is indicated by +. Details of the slopes are given in Table 3.2.

given in Table 3.2. Overall, the results show excellent linearity, and are quite close to the predicted beam angles: the vertical slope corresponds to a pointing angle of about 0.13 ± 0.13 degrees, while the off-vertical slopes correspond to a beam direction of about 11.1 ± 0.2 degrees. There is some deviation from linearity at the extreme outside antennas, which is due to the nature of the phasing hardware. These end antennas do not use the same electrical relay systems as the remainder of the array; hence, they have a small, unique phase-offset and are not included in the fits. We should also note that phasing is somewhat variable from day to day (less than $\pm 5^\circ$ variation in phase measurements of 0° – 180°), probably due to environmental effects; further examination could be warranted. However, we are satisfied with the beam directions, although making note of a possible biases in the vertical beam of about 0.1° which could fold $\sin(0.1^\circ) \approx 0.15\%$ of the eastward velocity into the vertical estimate.

Table 3.2: The phase slopes measured from the linear least-squares fits in Figure 3.2, given in units of degrees per antenna. The modelled off-vertical slopes are for a combination of 4° and 7° cables. The quoted error ranges are standard errors in the fitted slope, and end antenna values were not included in the fits (see text).

Beam Direction	Fitted Slope	Model
Vertical	-0.42 ± 0.42	0
West	34.17 ± 0.52	34.49
East	-35.25 ± 0.48	-34.49

That these phasing errors can be significant in estimates of the vertical velocity is an important point, and has implications for techniques which involve the measurement of phase (interferometry, for example). It is essential that any instrumental phase biases in a system which is used for such measurements are noted, and the appropriate corrections made, before ascribing any observed biases to geophysical effects (such as tilted scattering layers, for example). We shall discuss the effects of random phase variations further in the next section.

3.1.2.2 Array Polar Diagram and Phasing Errors

Random phase errors occur between coaxial-collinear rows, mainly due to less than perfect impedance matching at the feed point but also because of environmental effects, especially corrosive wear and moisture. The best beam-forming occurs when the signals at the antenna centre-feeds are all in phase (for a vertical beam), or are at regular intervals (for off-vertical beams). The centre-feeds of the Buckland Park VHF array are matched to a $50\ \Omega$ pure resistive impedance. The question here is, what effect do errors or random changes in this value have on the array polar diagram?

A major part of the CoCo array refurbishment involved recalibrating the dipole row impedance matching. A practical target was to bring the impedance at the feed point to within $5\ \Omega$ and 5° of the $(50\ \Omega, 0^\circ)$ pure resistive value: some connections had drifted over the years to more than $20\ \Omega$ and/or 30° reactance away from this ideal.

The array polar diagram was modelled by a simple aperture transform, treating the array as a set of point isotropic radiators at the location of each dipole antenna. Dipole rows run North-South, and are spaced $\lambda/2$ in the East-West direction. Cable losses

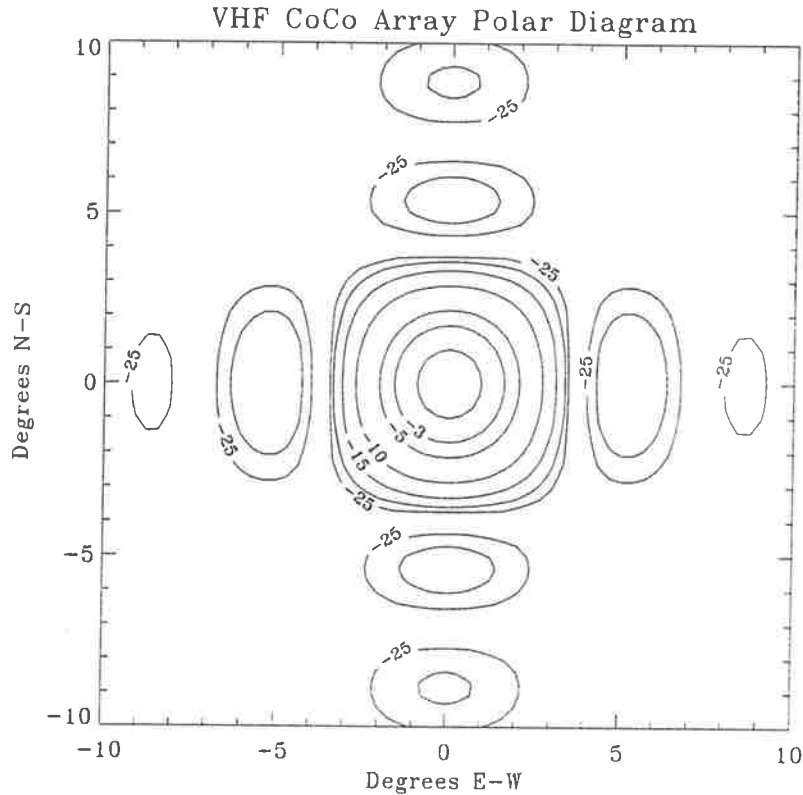


Figure 3.3: A contour diagram of the VHF CoCo antenna polar diagram. Contours are plotted at -1 dB, -3 dB, -5 dB, and every 5 dB thereafter to -25 dB. The half-power (-3 dB) half-widths are about 1.6° in the East-West plane, and about 1.7° in the North-South plane.

were included, and centre-feed phases were variable. Tilting of the beam was allowed via the introduction of incremental phasing between rows, exactly as occurs with the Buckland Park array in practice. The model code (in the IDL language) and a more complete description are given in Appendix B for reference. A careful comparison of the results for vertical and near-vertical beams displayed consistent results, so we only present results for the vertical beam here; however, our results are general for typical (small, near-vertical) beam tilts. A contour diagram of the ideal array, with no phase errors, is shown in Figure 3.3. The half-power half-width is about 1.6° in the East-West plane, and about 1.7° in the North-South plane (the E-W width is narrower than the N-S width because losses *along* the coaxial-collinear rows are more severe than losses in the transmission lines *to* each row). There is a slight displacement in the positions of the sidelobes due to the missing central CoCo row, which also accounts for the

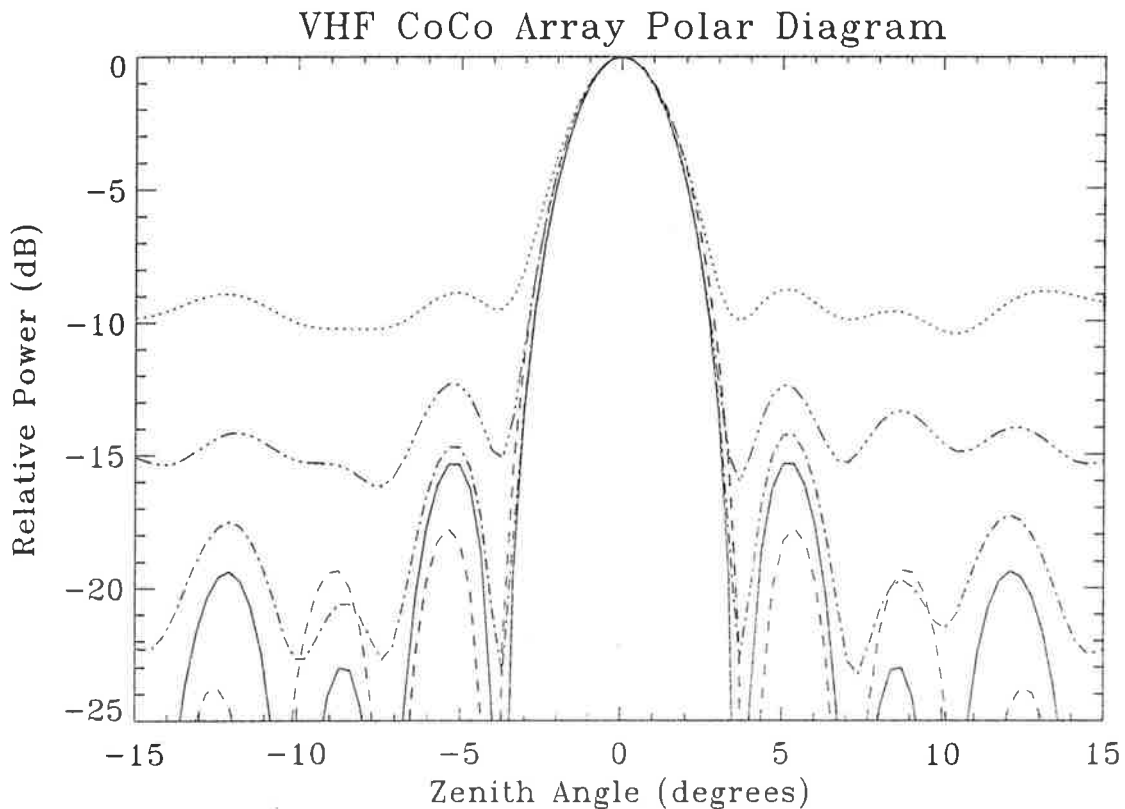


Figure 3.4: VHF CoCo antenna polar diagram cross-sections. Solid and dashed curves are ideal E-W and N-S sections respectively. Cross-sections in the E-W plane for Gaussian distributions of phase variations with half-widths 20° (chained), 45° (triple-chained), and 70° (dotted) are also shown. Gaussian magnitude variations of half-width 20% are also applied in the latter three curves.

extra suppression of the second sidelobe. The first sidelobe is suppressed about 15 dB compared to the main lobe.

The effect of random phases are displayed in Figure 3.4, which shows a number of cross-sections through the centre of the polar diagram. The N-S cross-section (dashed) obviously shows no effects of random phases (although note the good sidelobe suppression due to cable losses along the rows). The ideal E-W cross-section (solid) is shown for reference. The other E-W cross-sections are the averages of 100 realisations of array polar diagrams including random phases drawn from a Gaussian distributions centred on zero, with half-widths of 20° (chained), 45° (triple-chained), and 70° (dotted). These latter examples also include random impedance phasor magnitude variations drawn from a Gaussian distribution with a half-width of 20%, although the

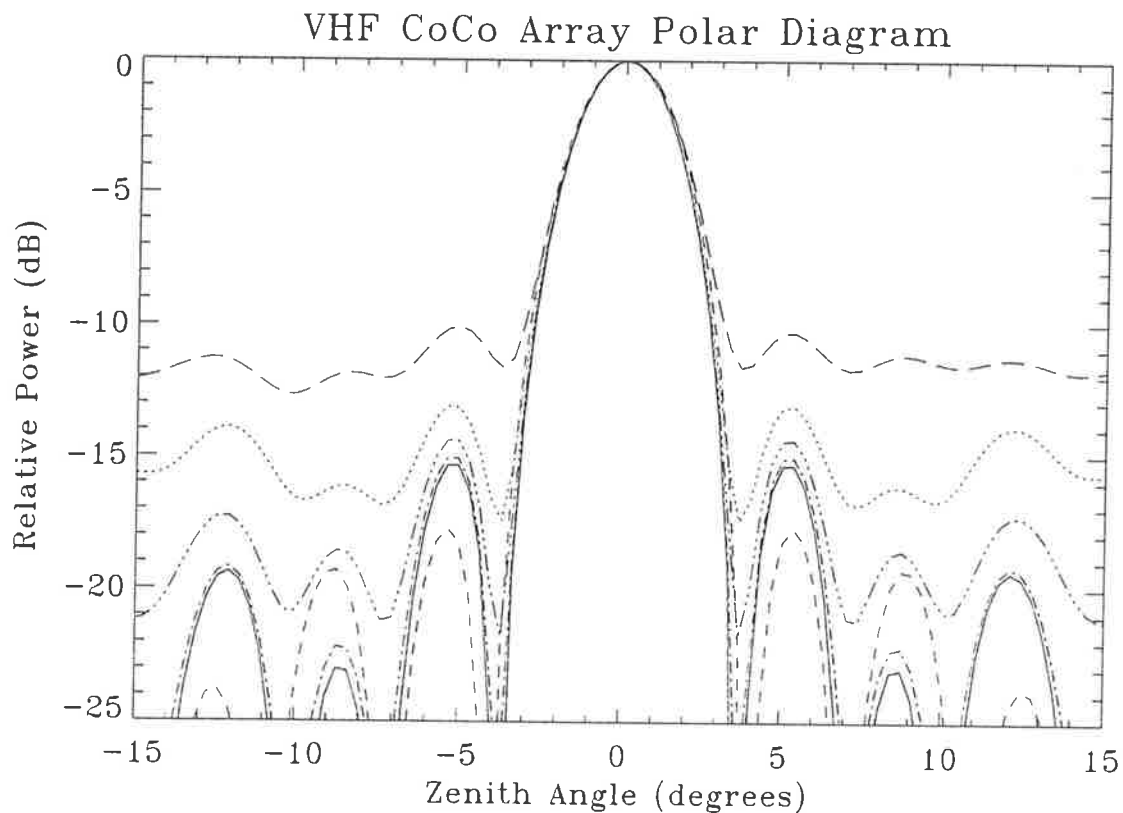


Figure 3.5: VHF CoCo antenna polar diagram cross-sections. Solid and short-dashed curves are ideal E-W and N-S sections respectively. Cross-sections in the E-W plane for Gaussian distributions of magnitude variations with half-widths 20% (chained), 50% (triple-chained), and 85% (dotted) are also shown. The long-dashed curve includes Gaussian-distributed phase variations with a half-width of 45° together with 85% half-width magnitude variations.

effect of magnitude variations is significantly smaller than that of phase variations. For example, it can be seen from Figure 3.5 that the inclusion of severe (85%) magnitude fluctuations to the 45° phase variations decreases the suppression of the first sidelobe by only 2 dB or so, and that of the background by less than 4 dB.

The effects are obvious—as the phase variation increases, the level of the sidelobes is increased relative to the main lobe, and the main lobe itself becomes slightly wider. However, the level of errors required to make a significant difference is large. Degrading the polar diagram affects directionality, the distribution of transmitted and received power around the sky. At 20 dB suppression, only 1% of transmitted power is returned from directions away from the main beam; at 10 dB suppression, the isotropic return increases to 10%. Obviously, this makes it harder to identify the origin of returns.

producing the same problems as discussed earlier with regards to broad beamwidths; it can also be considered as a loss in transmitted power. Eight years of minimal maintenance resulted in degradation of the array polar diagram bordering on significant to unacceptable, as far as accurate wind determination is concerned.

This simple model has also been compared with a full field analysis of the array designed for accurate meteor detection work, which displays quite similar results (M. A. Cervera, personal communication). We are satisfied with the present performance of the array, especially after retuning of the array elements and the instigation of a regular maintenance schedule.

3.2 MARWINsonde and Rawinsonde Systems

A collaborative observing campaign between the Department of Physics and Mathematical Physics at University of Adelaide, and the Centre for Dynamical Meteorology and Oceanography (CDMO) at Monash University, began in June 1991. Designated “SAFEx” (for the South Australian Fronts Experiment), it combined VHF wind profiler measurements at Buckland Park with co-located balloon-sonde launches. We discuss the SAFEx programme of observations in Appendix C, with example case studies in Chapters 6 and 7. The basis of operation of tracked balloon systems has already been discussed (Section 1.2.1); in this section, we present details of the precise systems utilised for data collection.

Together with the profiler observations, a Väisälä MARWIN rawinsonde system was operated on a campaign basis at the field site. The MARWIN is a portable system which uses standard RS80-15N radiosonde packages, identical to those used by the Australian Bureau of Meteorology, to measure temperature, pressure and humidity every two seconds (about 10 m in altitude), up to altitudes of about 25 km. The accuracy and precision of the observed parameters are summarised in Table 3.3. Software averaging in the radiosonde reception system produces a “cleaned” profile of the thermodynamic quantities, with spurious points rejected, at ten second (50 m) resolution.

Table 3.3: A summary of the sensor characteristics of the Vaisälä RS80-15N radiosonde package. “Accuracy” is determined by the standard deviation of the differences between repeated calibrations (*i.e.* repeatability). The sonde package weighs a total of approximately 205 g, with dimensions 55 mm x 147 mm x 90 mm.

	Pressure	Temperature	Humidity	Wind
Range	(3, 1060) hPa	(−90, +60) °C	(0, 100)% RH	(0, 180) ms ^{−1}
Resolution	0.1 mb	0.1 °C	1% RH	0.1 ms ^{−1} , 0.1°
Accuracy	±0.5 mb	±0.2 °C	±2% RH	0.5 ms ^{−1} (vectorial)

In addition, the MARWIN system has inbuilt windfinding capabilities: the balloon’s position is determined from the phase of received VLF signals transmissions over the Omega navigation network, which allows wind estimates to be calculated every ten seconds (about 50 m). The filtered wind profile is essentially a spline fit to 130–250 seconds of the raw data, after rejection of outliers [Väisälä, 1989]; accuracy is independent of range. The spline-fitting routines tend to smooth out the reported wind profiles, and moderate the effects of severe shear [Uddstrom, 1989]. The majority of Bureau radiosonde sites are equipped with standard radar-tracking of balloons for wind measurement, which has more severe limitations on observed range, resolution, and accuracy (around $\pm 2.5 \text{ ms}^{-1}$, increasing with range [Kessler *et al.*, 1985]).

Typically, the balloon takes about 90 minutes to reach 25 km, ascending at around 5 ms^{-1} . This sets an upper limit on the rate of launches, as only one sonde could be tracked at a time. An Intensive Observation Period (IOP) during SAFEx typically covered 36 hours with balloon flights every three hours or so. Launches from the Bureau-operated sites are made either 12- or 24-hourly.

3.3 Automatic Weather Station Observations

The final system that we should mention is the automatic weather station (AWS) which was installed at Buckland Park, again on a campaign basis. The AWS capabilities included the measurement of pressure, temperature, humidity, and wind speed and direction at the surface (more accurately, at 1.5 m above the surface), at one minute resolution (that is, the AWS package processes data and reports a filtered average

Table 3.4: A summary of the sensor characteristics for the devices mounted on the Buckland Park automatic weather station (installed by Monash University). The quoted resolution and accuracy are for the one-minute reported data, and are based on the factory-determined sensor quality.

	Pressure	Temperature	Humidity	Wind
Range	(600, 1050) hPa	(−50, +50) °C	(0, 100)% RH	(0.4, 75) ms ^{−1}
Resolution	0.3 hPa	0.2 °C	1% RH	1°, 0.1 ms ^{−1}
Accuracy	±0.5 hPa	±0.1 °C	±3% RH	±(3°, 0.1 ms ^{−1}) [< 10 ms ^{−1}] ±(3°, 2%) [10 to 75 ms ^{−1}]

every minute). Sensor details are displayed in Table 3.4. We should note that the usual meteorological requirement, that a sensor should not be located within $10h$ of an obstruction of size h , was not strictly observed due to space requirements; however, we did attempt to place the sensor in as open a location as was possible. In addition, we place little faith in the data from the AWS humidity sensor, which suffered from condensation problems. There is also the possibility of error in the AWS wind direction, due to inconsistencies in the setup procedure and a 8° difference between true and magnetic North. However, since we shall use AWS measurements as trend indicators rather than absolute measurements, this latter point is of little concern.

3.4 An Evaluation of Spaced Antenna Winds

We have already discussed the background to radar measurements of atmospheric wind profiles in Section 2.5, with reference to spaced antenna methods particularly in Section 2.5.2. Our aim here is discuss two experimental results from the observations taken during the tenure of this thesis: the effect of sampling time upon the Full Correlation Analysis (FCA) of Briggs [1984], and the performance of a “sliding” time average as applied to our wind comparisons. The wind comparisons presented here comprise the first such experiment carried out between a spaced antenna profiler and a large number of co-located radiosondes, minimising the effects of spatial variation.

3.4.1 The Effects of Sampling Time

As the SAFEx dataset increased in size over the two years of the experiment, a particularly disturbing trend was noticed in the spaced antenna wind retrievals: the performance around jet maxima was severely degraded (see, for example, Figure 6.13). Generally, the behaviour of the spaced antenna algorithm in regions of high wind (greater than about 30 ms^{-1}) was substandard, with extensive dropouts for computational reasons rather than low SNRs (compare Figure 6.19, for example, which shows perfectly normal SNRs around the region of spaced antenna wind estimation failure). The few results from such regions displayed an underestimation of the magnitude of the wind, as measured by MARWINsonde launches close in time and space to the radar estimates. However, the wind *direction* returned by the radar continued to be in excellent agreement with the sonde measurements. An example of this behaviour is shown in Figure 3.6(a), where the spaced antenna wind magnitude shows a distinct “levelling out” around 30 ms^{-1} . It is instructive to note that previous comparisons (Vincent *et al.* [1987], for example) have generally been made in conditions of low or moderate winds, mainly as a concession to spatial variability: in high winds, radiosondes can travel hundreds of kilometres laterally from their launch site while ascending to the mid-stratosphere, making precise comparison with ground-based sensors difficult. However, traces of the behaviour displayed in Figure 3.6 can also be seen in Figure 5 of Vincent *et al.* [1987] and Figure B.3 of Rajopadhyaya [1994].

To explain this result, it is worth taking a brief look at the assumptions of the FCA (see Briggs [1984] for the mathematical details). Essentially, the velocity of the atmosphere is extracted from the properties of the diffraction pattern (on the ground) caused by radio-wave scattering from refractive index irregularities. By correlating the time series of complex voltage measured at each spaced receiver, the pattern speed and time-evolution can be estimated directly from the time delays corresponding to the cross-correlation maxima; the atmospheric velocity can then be derived. In practice, the cross-correlation maximum and its position are determined by fitting a Gaussian or cubic spline to a few lags either side of the maximum. The lag time, of course, is determined by the sampling rate of the time series on the ground, and this is where

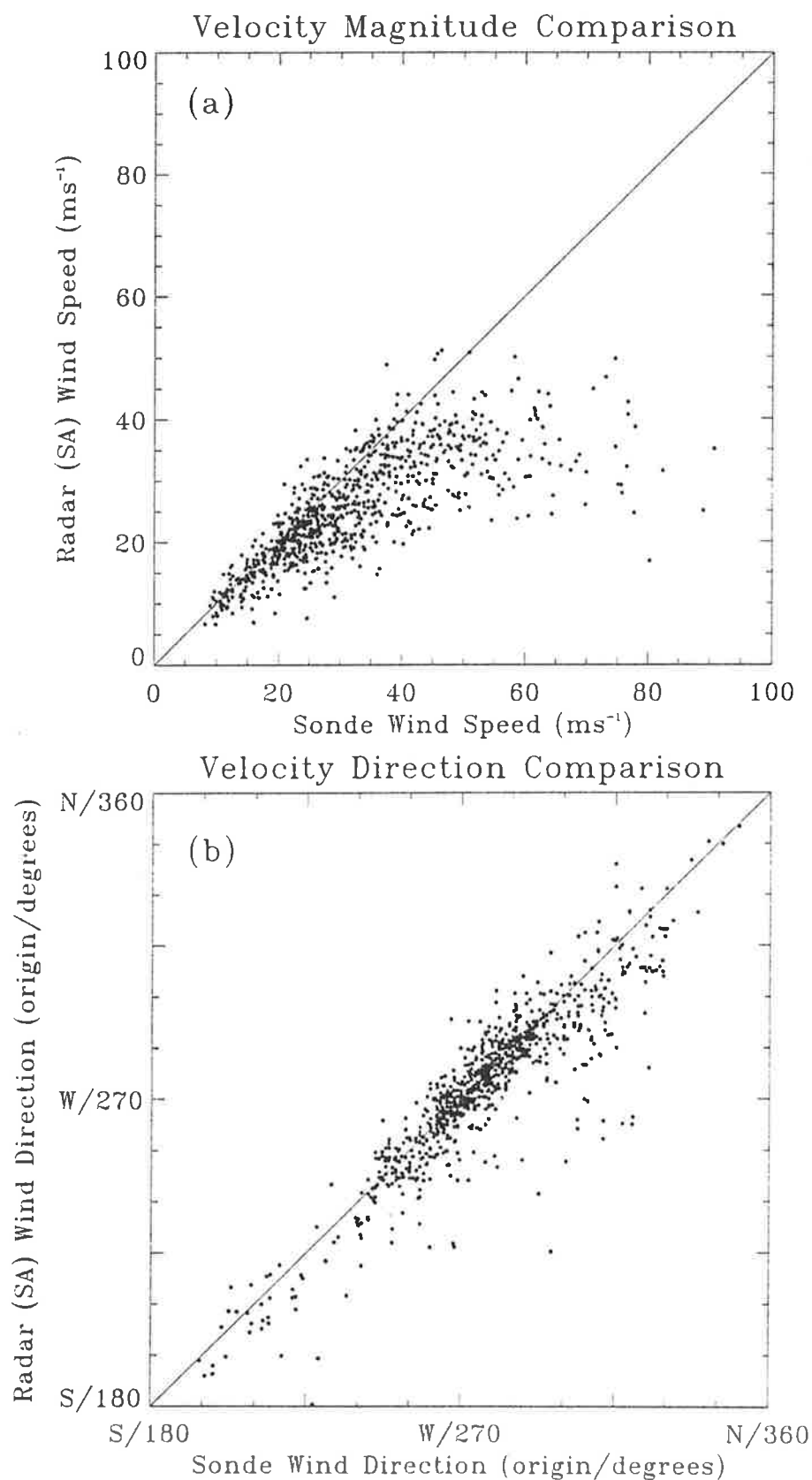


Figure 3.6: A comparison of spaced antenna winds with MARWINsonde retrievals, from the relevant IOPs in SAFEx Phase IV: (a) wind magnitude; and (b) wind direction. Each point represents an hour-long radar average centered on the time at which the sonde reach that height (a "sliding" average in height); there are 45 sondes and 26 radar range gates displayed in each plot, although missing sonde or radar retrievals reduce the total to 832 points.

problems can occur. If the sampling of the time series is too sparse, then a cross-correlation peak located near the first lag—or, in extreme cases, even closer to the zero-lag position—will be poorly determined, simply on grounds of resolution. A brief moment of reflection tells us that cross-correlations near zero-lag correspond to high velocities: if the time series at one antenna is nearly identical to that at another antenna, then the translational pattern velocity between those particular antennas must be relatively large. Note that the *direction* of the wind as determined by the FCA should not be affected to the same extent, as it is calculated from an arctangent of two velocity components where typically $|v/u| \ll 1$ (where $\mathbf{u} = (u, v, w)$). This is seen in Figure 3.6(b), where the correlation coefficient is 0.92.

As an example, consider two antennas which are 50 m apart, sampled every second. Assuming for simplicity that the pattern does not evolve in time, then if the pattern translates from one to the other at 50 ms^{-1} the time series recorded at each antenna will be identical, but separated by one lag; if the pattern translates significantly faster, then it will be very difficult to correlate the signals between the two antennas (barring spatio-temporal aliasing effects). The fewer lags one has available to make an estimate of the location of the cross-correlation peak, the less reliable the resultant estimate of atmospheric velocity.

Prompted by this initial treatment, *Holdsworth & Reid* [1995a] applied their FCA model to the question and demonstrated the problem theoretically. Using the parameters of the Buckland Park VHF spaced antenna system, they concluded that a sampling time of no more than 0.17 s should be employed to accurately estimate atmospheric velocities of 80 ms^{-1} or below; the maximum sampling time is inversely proportional to the maximum observable velocity. With the default 0.5 s sampling employed here, the maximum accurately-measurable velocity would be about 27 ms^{-1} , consistent with the observations displayed in Figure 3.6. Given that frontal jets near the tropopause can on occasion reach 100 ms^{-1} or more, a sampling time of 0.125 s or better would be a good figure to aim for, although one must always pay attention to the signal-to-noise tradeoff under faster sampling discussed earlier. We have made a preliminary set of measurements at the improved sampling rate, which has significantly improved performance at high wind velocities; a detailed experimental verification is the subject

of continuing and future examinations.

The sampling time problem seen in spaced-antenna/FCA techniques is a spatial analogue to the temporal limit described in Equations 2.10 and 2.11, due to the spacing of sensors on the ground and the use of spatio-temporal cross-correlations. The effect can severely curtail useful spaced antenna wind measurements, although it is gratifying to note that in most cases the FCA analysis returns a “no result” rather than an erroneous wind estimate. For example, the extended region of missing data in Figure 6.13 (a two-hour average, where an average is only recorded if good data is available for at least 25% of the averaging interval) indicates that over 75% of the estimates made there simply failed outright to pass the rigorous FCA rejection criteria (as described by *Briggs* [1984] and others). Although accurate returns would be the best result, a failed estimate is far better than an incorrect one. Future experimenters would be well advised to take account of our experiences when considering appropriate sampling rates.

3.4.2 Comparison with MARWINsonde Winds

Before wind profiler data derived from any method (Doppler beam swinging, spaced antenna, interferometry, *etc.*) can be applied routinely in operational forecasting, it is necessary to compare the results with other—standard and accepted—techniques such as rawinsonde retrievals. The discussion of the history of such measurements in Section 2.5 highlights two concerns: a lack of extensive spaced antenna winds comparisons in general; and a dearth of comparisons between nearby or co-located systems for any radar technique. Differences between radar and rawinsonde profiles can be attributed to two main sources: errors and variances inherent to each particular system; and variations between what each system actually measures. For example, the radar wind profile is over a single location in space, with all ranges measured simultaneously, while the sonde data at each height are recorded at different times over different locations due to advection of the sensor with the wind and a relatively slow ascent rate. Hence, spatial and temporal variations in the wind field would be expected to contribute to differences between the wind profiles measured by each system.

A first attempt at correcting for spatial variations is to use co-located systems, reducing the initial difference to a minimum, although the variation will increase with time (and height). Taken in isolation, this assumes that the wind field does not evolve in time; rather, that it is “frozen” with some spatial distribution ($\mathbf{u} = \mathbf{u}(x, y, z)$). If one has some knowledge of the wind field around the launch site (from numerical forecasting models or assimilation analyses, for example) then spatial gradients of the wind can be calculated, which would allow for further correction. However, horizontal wind gradients are generally quite small, around $0.001\text{--}0.02\text{ ms}^{-1}\text{ km}^{-1}$, so that variations over the order of 100 km amount to just 1 ms^{-1} or so [Steinhagen *et al.*, 1994]. As seen in Section 2.5 and Table 3.3, and again below, this large-scale variation is usually swamped by inherent system errors and short-term wind variability. Indeed, the small spatial gradients on the β -mesoscale are usually difficult to separate from those associated with temporal evolution, even when multi-station observations are available.

Alternatively, one could assume that the wind field is fixed over a two-dimensional slice of the atmosphere, but can vary with time; that is, the wind is purely a function of height and time ($\mathbf{u} = \mathbf{u}(z, t)$). The usual method employed for temporal averaging is to simply average the radar data for, say, an hour (being a reasonable estimate of the time for the balloon to reach the maximum radar range) at each height, starting from the time of balloon launch [Weber & Wuertz, 1990]; equivalently, one could centre the averaging interval on the middle of the balloon ascent time. However, neither of these approaches takes into account the evolution of the wind field with time. Assuming for the moment that the balloon ascends vertically, with no horizontal displacement, then a more representative radar parameter average at a given altitude is one centred on the actual time that the balloon reached that height. We label this as a “sliding” average. For a balloon launched at some time t_0 , ascending at a speed v_b , and reaching some height z at time $t' = t_0 + z/v_b$, the sliding average of a wind component ($\mathbf{u}_x = u$, for example) over an hour at that altitude is:

$$\overline{u(z)}|_{t=t'} = \left(\frac{1}{m}\right) \sum_{t=t'-30}^{t'+30} u(z, t) \quad (3.2)$$

(altitude in metres, time in minutes, u in ms^{-1} and m radar estimates per hour). The

standard average would have summation limits of $[t_0, t_0 + 60]$ (starting from the time of the balloon launch), while the “centered” average would sum between $[t_{0.5} \pm 30]$ where $t_{0.5}$ is halfway through the flight (if the flight finishes at time t_f , then $t_{0.5} = (t_0 + t_f)/2$).

Our comparisons are limited because of the sampling time problem and our resultant inability to compare velocities greater than about 30 ms^{-1} . However, our results are extensible to larger velocities (when appropriate sampling is employed) since finer-scale sampling does not affect the validity of the method.

Figure 3.7 shows a comparison of spaced antenna and MARWINsonde winds as a time series at 3 km altitude for about four days during the SAFEx campaign IOP IV/3 (Appendix C); note that time runs from right to left. The variability measured by the profiler is relatively high during the last three days, as a consequence of the passage of a mesoscale convective system (see Chapter 6). However, rather than being a disadvantage, this reflects the behaviour of the atmosphere on scales which the conventional radiosonde measurements cannot match. Since the sondes make essentially point measurements in time, there is a tendency for the sonde wind magnitude measurement to be made away from the mean profiler record. In essence, if the wind happens to be anomalously high or low when the ascent is made, then the sonde may report a “non-mean” velocity. Note that the sonde measurements fall nearly without exception into the region defined by the radar-measured variability. The resolution of the profiler also highlights some features which even these relatively high-resolution sonde measurements miss: the velocity minimum near 0900 UTC 30 August, for example. Routine twelve-hourly sonde flights certainly present an artificially smoothed picture of atmospheric wind variability. However, the agreement between the sonde and radar is generally excellent, given the bounds of inherent variability.

An example altitude profile comparison is displayed in Figure 3.8, which again displays excellent agreement. Note the effects of the sampling time problem at the highest altitudes, where the radar begins to severely underestimate the actual atmospheric velocity above about 30 ms^{-1} . The radar data has been averaged by the sliding method here; extensive comparisons with the centered method reveals only slight visual differences for the most part, although the sliding average appears to bring the magnitudes of wind extrema into better agreement in many cases. It would be fair to

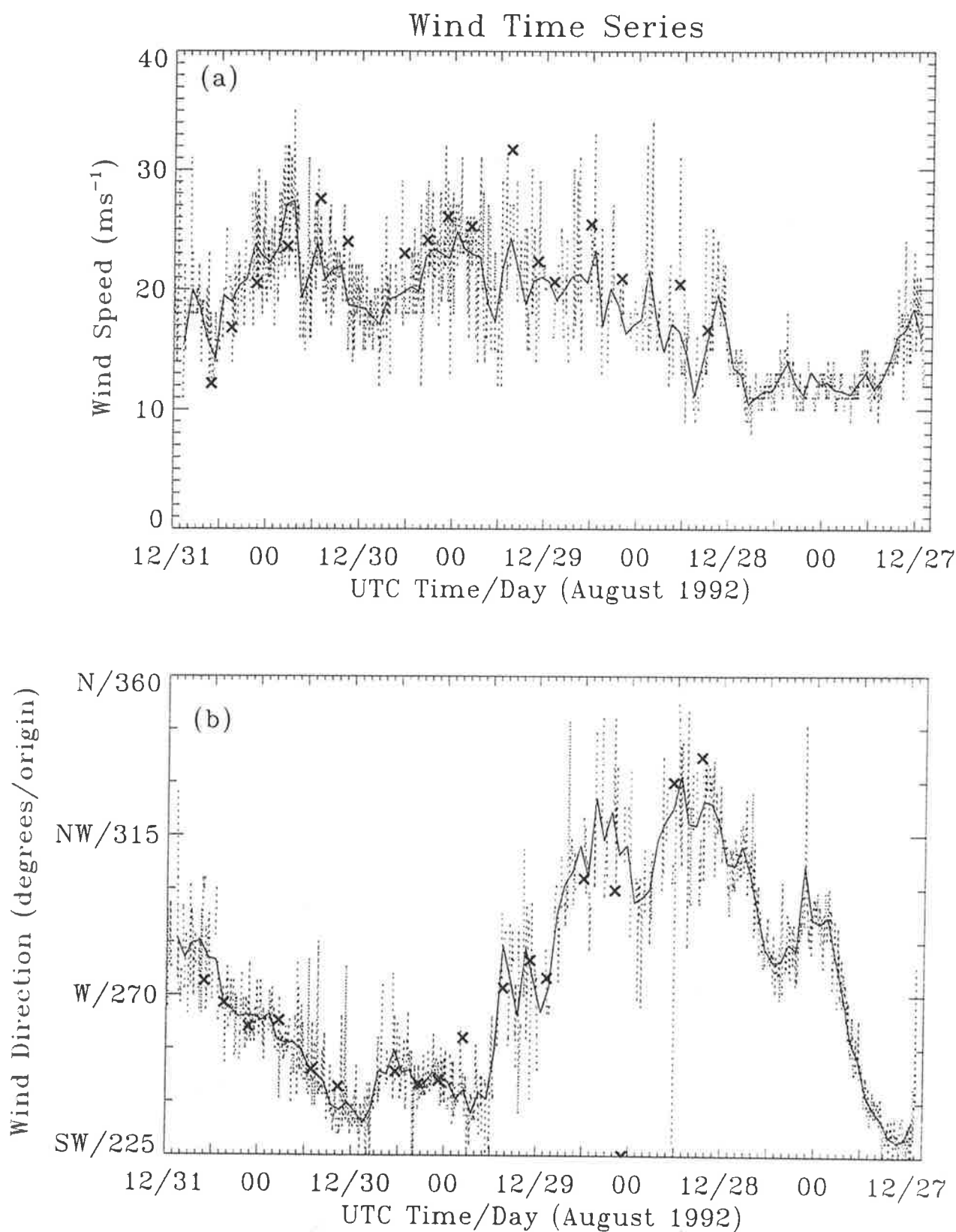


Figure 3.7: Time series of (a) wind speed and (b) wind direction at 3 km altitude from SAFEx IOP IV/3. Sonde measurements are marked by $\times$'s, raw spaced antenna winds by dotted lines and two-hourly average spaced antenna winds by solid lines.

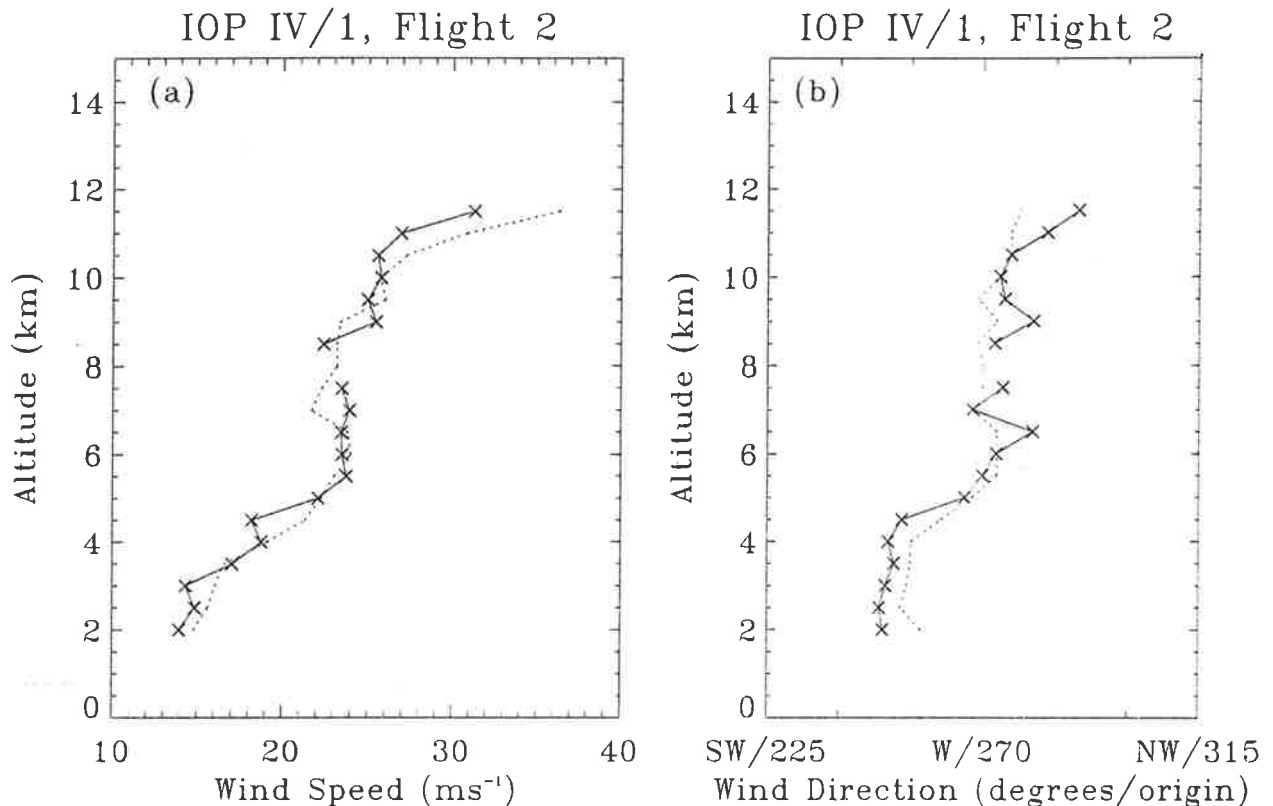


Figure 3.8: Altitude profiles of (a) wind speed and (b) wind direction for profiler (solid line, $\times$'s) and MARWINsonde (dotted line), from SAFEx IOP IV/1 (flight 2). The profiler data are one hour sliding averages; error bars are not shown for clarity, but the natural variability $\sigma_{|u|} \sim 4 \text{ ms}^{-1}$ and $\sigma_{\theta} \sim 7^\circ$ (see Figure 3.7).

say that the natural variability exceeds the difference between the sliding and centered averaging methods, although a statistical treatment is helpful in order to place some limits on the degree of difference.

If we restrict ourselves to cases where the atmospheric velocity is less than 25 ms^{-1} —which, in effect, limits our comparisons to below about 6 km altitude—then over the 98 flights where coincident sonde and spaced antenna wind data are available, we arrive at the results presented in Table 3.5. The size of the bias, standard deviation, and RMS fluctuations are all comparable with the previous results cited earlier concerning radar-sonde intercomparisons. The differences between the sliding and centered averaging methods are not strictly statistically significant (the degree of significance is less than 90%), although there is an indication of the observed small improvement. A detailed examination of individual profiles reveals that there are a number of cases

Table 3.5: Spaced antenna versus MARWINsonde wind comparison statistics: the bias, or mean difference; the standard deviation σ associated with the bias; and the total RMS fluctuations over the entire dataset. Wind magnitudes are in ms^{-1} , directions quoted in degrees (wind origin); all differences are radar winds subtracted from sonde winds. [C] and [S] refer to “centered” and “sliding” averages respectively. Only data with wind magnitudes less than 25 ms^{-1} have been used, to avoid the sampling time problem.

		Bias	σ	RMS
Magnitude:	[C]	1.0	1.7	4.3
	[S]	0.8	1.6	3.5
Direction:	[C]	4.3	7.8	10.0
	[S]	3.7	6.8	9.5

where the sliding average corrects an individual point or a few points in a given profile. However, the natural atmospheric variability seen in the profiler measurements, with a variance of about $\pm 4 \text{ ms}^{-1}$, tends to dominate the comparisons.

As mentioned earlier, the radar and sonde measure different aspects of the same atmospheric wind field. Taking these differences in temporal and spatial resolution into account, the overall agreement between the two systems is excellent. The sliding average appears to make a small improvement by reducing the effects of the temporal separation of measurements at a given altitude level, although a larger sample size would be required to demonstrate statistical significance to better than the 95% level. Since the variability due to spatial gradients is also quite small, we are led to the conclusion that natural atmospheric variability (due to ubiquitous short-period gravity waves, for example) is the major contributor to the observed differences, together with any residual spatial variability.

3.5 Vertical Velocity

The understanding of vertical circulation in the atmosphere is of fundamental importance, with applications to a wide variety of fields including meteorological forecasting, dynamics, and the vertical transport of trace constituents. In many ways, vertical air motion is the single most important variable for many forecasting applications, and

much of synoptic meteorology is involved with the prediction of vertical velocity. For example, many forecasting “rules of thumb” (such as those involving convergent and divergent processes) have their roots in the dynamical equations for vertical motion. The development of precipitation (within frontal structures, for example) can be related to the location of regions of ascending air. The dynamics of tropospheric vertical transport (due to convective processes) and cross-tropopause mixing can be related to the mean global circulation.

At present, observed vertical velocities are not incorporated into numerical weather prediction (NWP) models; rather, the vertical velocity is a derived field. We can expect model performances to improve as profiler measurements become more widely available to, and accepted by, the meteorological community. Observations of vertical velocity around mesoscale events, such as cold fronts and baroclinic storm systems, provide model verification and an impetus to improvements therein; better pictures of mesoscale events allow for a more accurate prediction of system behaviour. Additionally, real-time monitoring of the vertical velocity field can improve severe weather warnings, particularly in the critical zero to six-hour timeframe. Other parameters, such as diabatic heating rates, can be directly computed rather than indirectly inferred with the inclusion of observed vertical velocity fields.

Wind profiling radars present a unique way to measure the vertical velocity, but to place the observations into context it is worth considering the standard methods of estimating vertical velocity. We then examine another, semi-direct measurement of vertical velocity—via balloon ascent rates—before discussing some characteristics of radar estimates.

3.5.1 Analyses

The standard meteorological methods for determining vertical velocity involve large-scale, indirect estimates from synoptic analyses; some of the common methods are outlined below. Intercomparisons began half a century ago (*e.g.* Panofsky [1946]) and are a continuing topic of research. Each indirect method has inherent errors and uncertainties due to assumptions (such as adiabatic flow) in the derivation, and inaccuracies

or inadequacies of the input data. *Belt & Fuelberg* [1982] found that vertical velocities calculated with subsynoptic scale data were as much as 6 cm s^{-1} larger than those calculated with synoptic scale data in the same conditions. This smoothing of vertical velocity should be borne in mind when comparing vertical velocities from analyses with single-station data (see, for example, *Nastrom et al.* [1985a], *Larsen et al.* [1988], and *Nastrom et al.* [1994]).

The simplest large-scale technique is the divergence method, derived from the equation of continuity assuming incompressible flow:

$$\frac{\partial w}{\partial z} = -(\nabla \cdot \mathbf{u}_h) = -\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) \quad (3.3)$$

where $\mathbf{u}_h = (u, v)$ is the horizontal wind. Direct calculations show that derived values of w from this technique are very sensitive to small changes in the distribution of wind speed and direction; the latter is not known accurately enough to make such direct calculations worthwhile. Instead, practical applications infer w indirectly from vorticity: essentially, one uses vorticity to calculate $(\nabla \cdot \mathbf{u}_h)$ and then derives $\partial w / \partial z$ from the divergence equation. The theory is discussed in most texts on meteorology (see, for example, *McIntosh & Thom* [1981], although it can become algebraically complicated. The quasi-geostrophic omega equation, for example, is derived from the vorticity equation and the first law of thermodynamics assuming the flow is isentropic and quasi-geostrophic. Other assumptions and a derivation are given by *Holton* [1992]. *Nastrom et al.* [1985a] use the following version of the equation:

$$\left(\sigma_c \nabla^2 + f_{c0}^2 \frac{\partial^2}{\partial p^2}\right) \omega = f_{c0} \frac{\partial}{\partial p} [\mathbf{u}_h \cdot \nabla (f_{c0}^{-1} \nabla^2 \Phi + f_c)] - \nabla^2 [\mathbf{u}_h \cdot \nabla (\partial \Phi / \partial p)] \quad (3.4)$$

where σ_c is the static stability parameter constant at each pressure level p , f_c is the Coriolis parameter, and Φ is the geopotential height (see also *Tapp & Oliver* [1989]).

The adiabatic method is based on the first law of thermodynamics, assuming isentropic flow. The vertical velocity ω in pressure coordinates is given by:

$$\omega = -(\partial \theta / \partial t + \mathbf{u}_h \cdot \nabla \theta) / (\partial \theta / \partial p) \quad (3.5)$$

where θ is potential temperature and p is pressure. The resultant values of ω can be converted to vertical velocity w in height coordinates through:

$$w = -(RT/pg) \omega \quad (3.6)$$

where R is the gas constant, T is temperature and g is the acceleration due to gravity.

The kinematic method is derived from upward integration of the equation of continuity:

$$\omega = \omega_0 + \sum_i (\nabla \cdot \mathbf{u}_{hi}) \Delta p_i \quad (3.7)$$

where ω_0 is evaluated from appropriate boundary conditions: as a simple example, we could take the vertical velocity at ground level to be zero, while more complicated methods make use of orographic and frictional effects. Surface divergence provides an important contribution to accurate kinematically-computed vertical velocity estimates, suggesting that the resolution of surface sensors is a key consideration.

3.5.2 Tracked Balloon Estimates

Since we have access to high-resolution balloon data, the possibility of extracting vertical velocities from the balloon ascent rates deserves examination. If realistic estimates of vertical velocity can be made from the sonde data, they would prove a valuable source of information for comparison with, and validation of, the profiler estimates.

The pioneering work of *Corby* [1957] investigates balloon ascent rate variations via the recorded change in pressure with time (the omega velocity, $\omega = dp/dt$) converted to vertical velocity, $w = dh/dt$. Comparison with radar tracking data shows the existence of statistically significant oscillations, associated with orographic sources. *Reid* [1972] perform essentially the same experiment, although with a new generation of radiosonde technology. Further applications of the technique can be found in the work of *Lalas & Einaudi* [1980], *Kitchen & Shutts* [1990], *Shutts & Broad* [1993] and *Shutts et al.* [1994].

Massman [1978] considered the neutral buoyant oscillations of a spherical, constant-volume balloon floating in the upper troposphere or lower stratosphere, and found that the balloon's response is more strongly influenced by time variations in the equilibrium density surface than by variations in atmospheric vertical wind speeds: that is, the balloon acts as a true isopycnic tracer. *Nastrom* [1980] placed limits on this conclusion by considering the action of waves on the balloon in regions of different static stability,

finding that the behaviour varied from “true isopycnic” as conditions departed from long-period waves in regions of high static stability.

How can we extract the vertical motion of the air from records of the balloon ascent rate? If we make the reasonable assumption that the air motion is additively separable from the total balloon motion (which is, in essence, the basis of tracked-balloon winds!) then, if we can calculate the balloon motion through an atmosphere at rest, we can determine the vertical velocity of the atmosphere from the total ascent rate. Note that we are assuming that the vertical motion of the sensor is equal to the vertical motion of the air; that is, that there is no air motion relative to the sensor. Without the type of experiment performed by *Sidi et al.* [1988], using multi-axial anemometers, we are restricted to assuming that the balloon does in fact travel with the mean wind. When the motion of the balloon is dominated by other effects (the possibility of which is discussed below), estimates of vertical velocity are bound to be compromised.

Consider a balloon filled with helium (or some equivalent balloon gas) of initial filled radius R_0 , ascending through an atmosphere with sample profiles of pressure $p(z)$ and temperature $T(z)$. The balloon’s radius and volume as a function of height are $R(z)$ and $V(z)$ respectively. Apply the ideal gas law to the gas inside the balloon:

$$pV/T = \text{constant} = \gamma \quad (3.8)$$

and hence:

$$\gamma = \left(\frac{4}{3} \pi R_0^3 \right) \frac{p_0}{T_0} \quad (3.9)$$

The balloon expands as it ascends, such that:

$$V(z) = \frac{4}{3} \pi R^3 = \gamma T/p \quad (3.10)$$

or, alternatively:

$$R(z) = R_0 \left(\frac{p_0}{T_0} \frac{T}{p} \right)^{1/3} \quad (3.11)$$

with $R = R(z)$, $T = T(z)$, and $p = p(z)$. Note that we are assuming equilibrium between conditions internal and external to the balloon; later, we discuss a more realistic set of parameters. Note that the filling of the balloon is not a precise activity, and estimates of initial radius R_0 based on observed lift, approximate filling-guide size,

and the number of balloons per gas cylinder give values between about 0.55 and 0.70 m in practice. Since balloon volume, and hence mass and lift, are highly dependent on initial radius, there are clearly limits on ascent rate predictions.

A typical balloon flight is displayed in Figure 3.9, with background pressure and temperature profiles, the observed ascent rate profile, and the estimated balloon radius as a function of altitude. The errors in measured ascent rate increase exponentially with altitude, since the absolute pressure error is constant with height, and balloon altitudes are derived from the hydrostatic equation; numerically, the errors in ascent rate are about $\pm 0.1 \text{ ms}^{-1}$ at ground level, $\pm 0.5 \text{ ms}^{-1}$ at 15 km, and $\pm 1.0 \text{ ms}^{-1}$ at 20 km. This sets some limits on the scale of oscillations we can observe, which are similar to the limitations for other techniques. Full details of the MARWINsonde system parameters are given in Section 3.2 and Table 3.3: we do not show error bars in the interests of clarity. The measurement limits should be borne in mind, however, and we shall draw attention to them where appropriate. For example, random errors in the balloon radius due to pressure and temperature measurement errors are less than 1% over the entire flight; differences in the initial radius simply translate the profile.

Next, consider the forces acting on the balloon, as depicted in Figure 3.10. The buoyancy contribution is given by:

$$F_B(z) = \rho_{\text{air}}(z)g(z)V(z) \quad (3.12)$$

where the air density ρ_{air} is given by:

$$\rho_{\text{air}}(z) = \frac{\bar{m}}{\tilde{R}} \frac{p(z)}{T(z)} \quad (3.13)$$

where $\tilde{R} = 8.3143 \text{ J K}^{-1} \text{ mol}^{-1}$ is the gas constant per mole, $\bar{m} = 28.8 \times 10^{-3} \text{ kg}$ is the molecular weight of air, and gravity varies above an Earth of radius r_E by:

$$g(z) = g_0 / (1 + z/r_E)^2 \quad (3.14)$$

We can neglect the errors in g with altitude as insignificant compared to the errors in determining R_0 , so that:

$$F_B = \frac{\gamma \bar{m} g_0}{\tilde{R}} = \text{constant} \quad (3.15)$$

as expected, since the mass of helium inside the balloon is conserved.

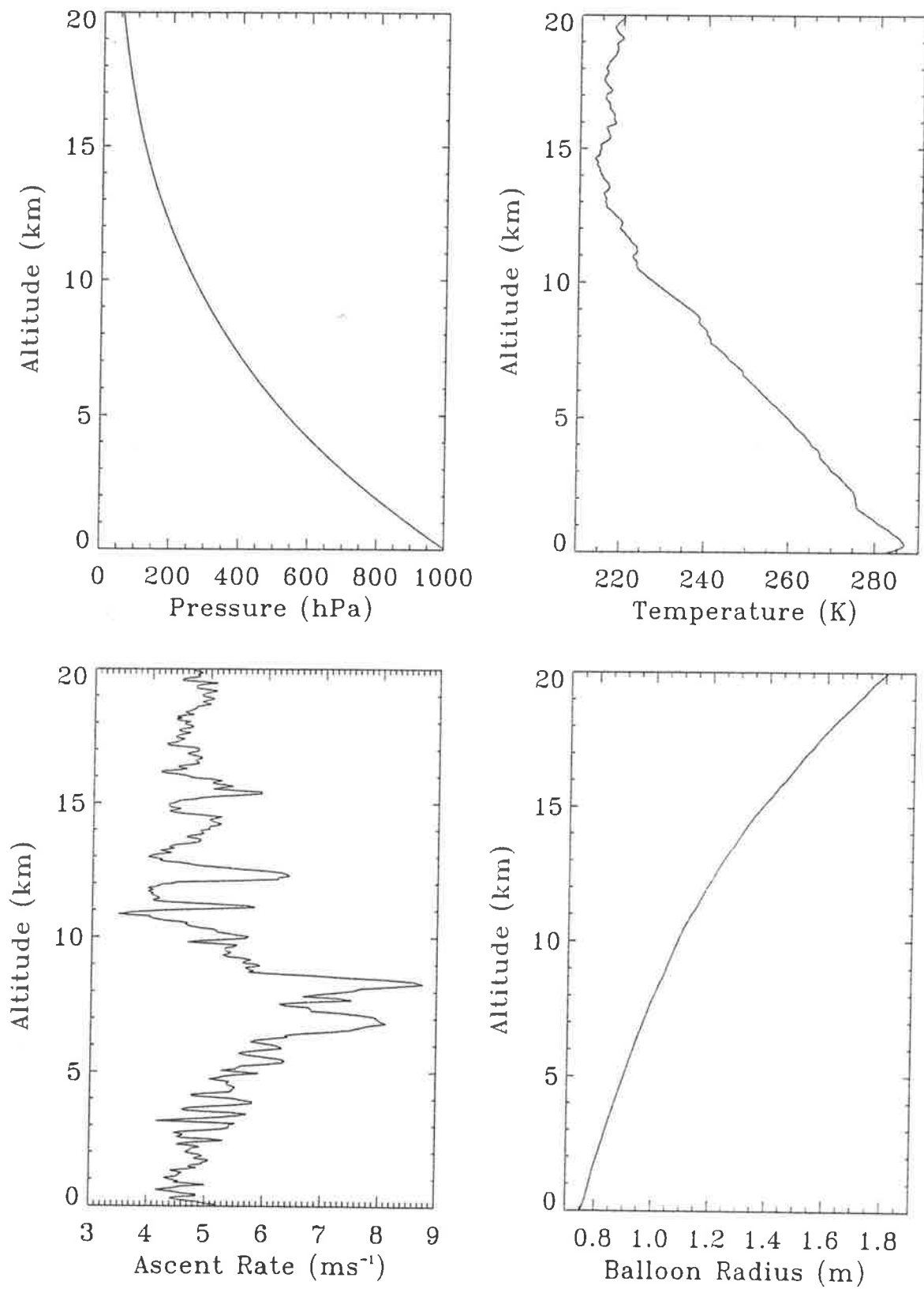


Figure 3.9: Typical temperature, pressure, ascent rate and calculated balloon radius profiles. The data is taken from SAFEx IOP IV/3, flight 1 (0019 LST 29 August 1992). Refer to the text for a discussion of measurement errors.

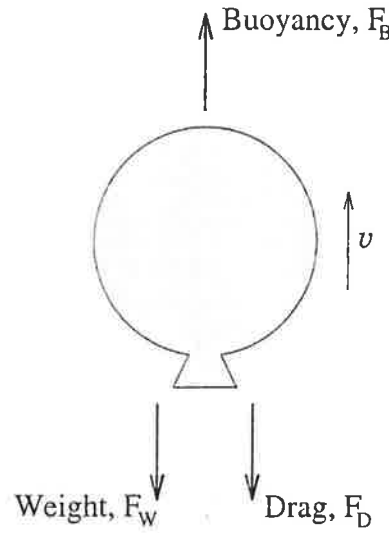


Figure 3.10: Forces acting on a freely-ascending balloon.

The weight force is similarly constant, since:

$$F_W = m_{\text{total}}g \quad (3.16)$$

$$= \rho_{\text{He}} V_{\text{balloon}}g + m_{\text{payload}}g \quad (3.17)$$

$$= \text{constant} \quad (3.18)$$

Hence, in the absence of drag, the balloon experiences a constant acceleration, directed positive-upwards, due to a force:

$$F = F_B - F_W \quad (3.19)$$

Comparing to the classical case of a falling raindrop, and taking note of our observational expectations of near-zero net force (as the balloon ascends at a near-constant velocity), we expect the drag force F_D to balance. That is:

$$ma_{\text{total}} = F_B - F_W - F_D = 0 \quad (3.20)$$

In other words, we expect a terminal velocity, v_T .

Any formulation of drag is highly dependent on the nature of the fluid flow, which can be parameterised by the Reynolds number:

$$R_e = vL\rho/\eta_v \quad (3.21)$$

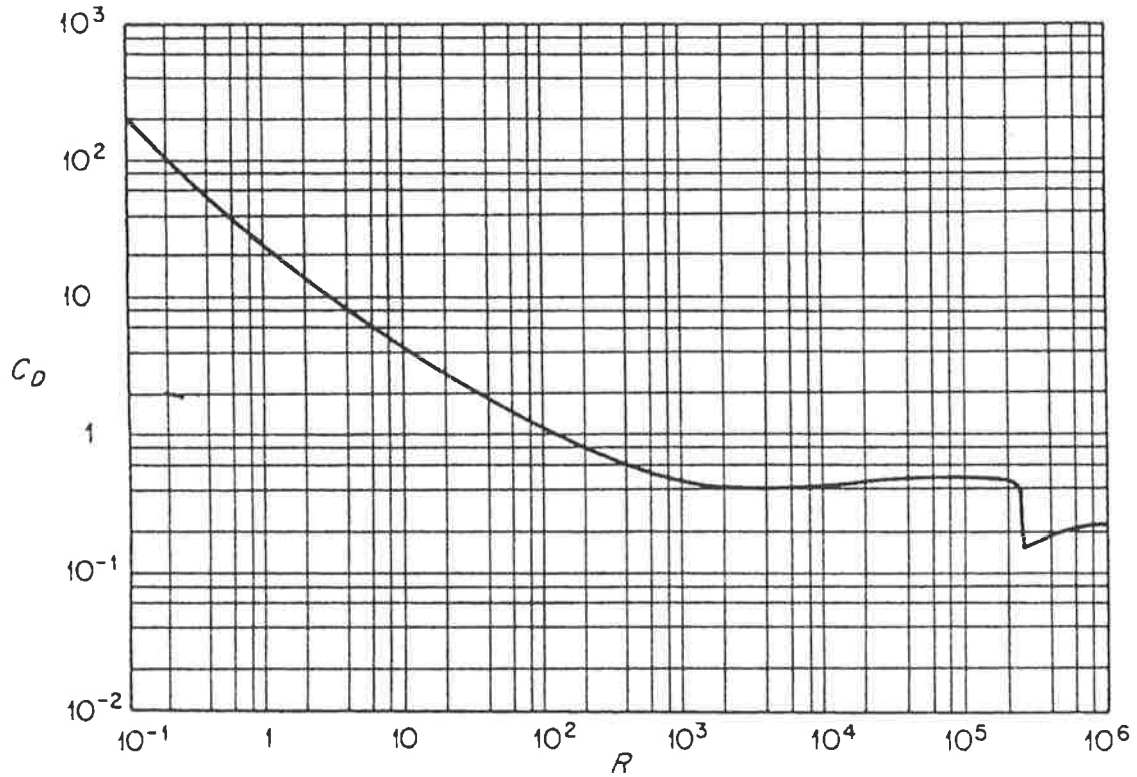


Figure 3.11: Drag coefficient of a sphere as a function of Reynolds number (from *Yih* [1969]).

where in our case $v \sim 5 \text{ ms}^{-1}$, $L \sim 1 \text{ m}$, $\rho \sim 0.1 \text{ to } 1 \text{ kg m}^{-3}$, and $\eta_v \sim 1.5 \times 10^{-5} \text{ kg m}^{-1} \text{ s}^{-1}$ are typical velocity, length, density and viscosity scales; hence, $R_e \sim 10^5$. For a sphere, this is in a transition region with regards to flow separation; the drag coefficient C_D varies sharply, as shown in Figure 3.11. The drag force can be represented by:

$$F_D \approx C_D \rho_{\text{air}} v^2 \pi R^2 / 2 \quad (3.22)$$

and C_D drops from ~ 0.5 to ~ 0.2 as R_e increases over $\sim 2.5 \times 10^5$.

It is reasonable to expect $v \rightarrow v_T$ rapidly, since the environment is typically low-resistance, and the drag force goes as the square of the airspeed. This is borne out by our observations, which do not show any slow tendency towards a limiting case. Hence, using Equation 3.13:

$$v_T \approx \sqrt{\left[\frac{2\tilde{R}(F_B - W)}{C_D \bar{m} \pi R^2} \right] \frac{T(z)}{p(z)}} \quad (3.23)$$

Note that $R(z) \propto (T(z)/p(z))^{1/3}$, so that $v_T \sim (p/T)^{-1/6}$ (using Equation 3.11). This is strongly dependent on the pressure term, as seen in Figure 3.12, which also demonstrates the small effect of temperature perturbations. The predicted ascent rates vary strongly with the initial radius of the balloon, and the limits on measuring R_0 discussed previously are obviously important in setting the “base” ascent rate. However, since R_0 cannot be height-dependent, it does not contribute to observed fluctuations in ascent rate.

Also, note that $v_T \propto C_D^{-1/2}$, which is a far stronger dependency than either pressure or temperature. Figure 3.13 illustrates that even small changes in C_D produce large changes in terminal velocity and, as mentioned above, significant changes in C_D are likely due to changes in Reynolds number. The numbers quoted above are for a sphere, which leads to our first extension: how does the shape of the balloon affect C_D ? A few cases are displayed in Figure 3.14; in practice, we would expect some flattening at the top of the balloon due to resistance at the balloon-air interface, together with elongation in the wake due to the weight of the sonde package. Indeed, one should not discount the possibility of the balloon undergoing a flexing-type oscillatory motion about one or more axes; such a flexing would change the balloon’s drag coefficient markedly between extrema, likely resulting in oscillatory ascent rate variations. An in-depth investigation of the changes in balloon shape during ascent is beyond the scope of this thesis, but we can make some qualitative observations.

Figure 3.15 shows the variations in drag coefficient which would be required to produce the observed ascent rate profile from a SAFEx balloon launch, calculated by solving Equation 3.23 for C_D assuming that the observed ascent rate is equal to v_T at every point. It is interesting to note that the spread of values, even with the assumptions we have made, fall almost entirely between the approximate theoretical limits of 0.2 and 1.2 (based on the shape and flow considerations presented above). Other balloon flights show similar behaviour, which implies that variations in drag coefficient (associated with shape fluctuations and changes in the nature of the flow) are dominant. At the very least, the ascent rate fluctuations are consistent with possible changes in the drag coefficient. This is referred to by *Reid* [1972], who speculates that “...the drag coefficient can change rapidly enough to markedly change the slope of

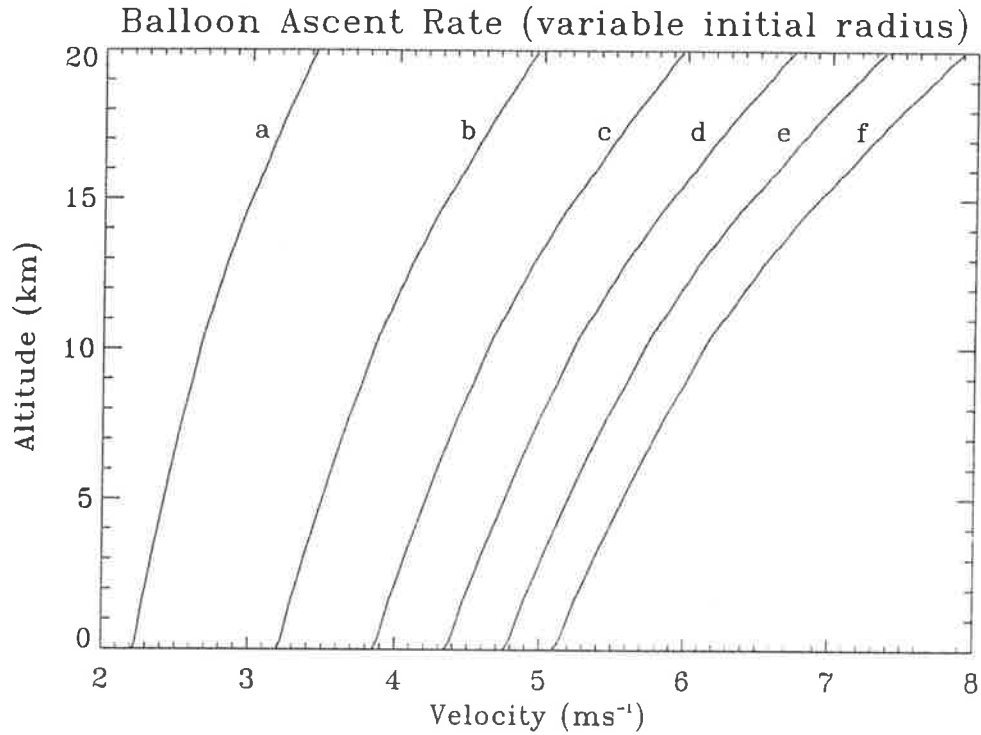


Figure 3.12: Terminal velocity for a fixed drag coefficient ($C_D = 0.5$) and variable initial radii: curves (a)–(f) correspond to initial radii of 50, 55, 60, 65, 70 and 75 cm respectively.

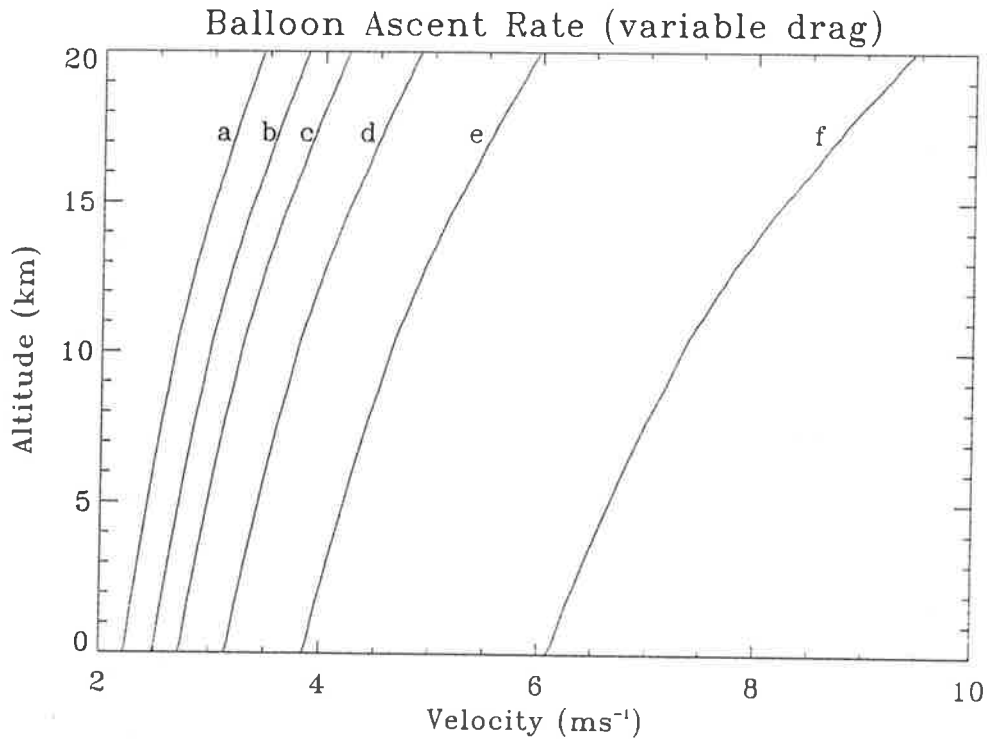


Figure 3.13: Terminal velocity for a fixed initial radius ($R_0 = 60$ cm) and variable drag coefficients: curves (a)–(f) correspond to height-constant drag coefficients of 1.5, 1.2, 1.0, 0.75, 0.5 and 0.2 respectively.

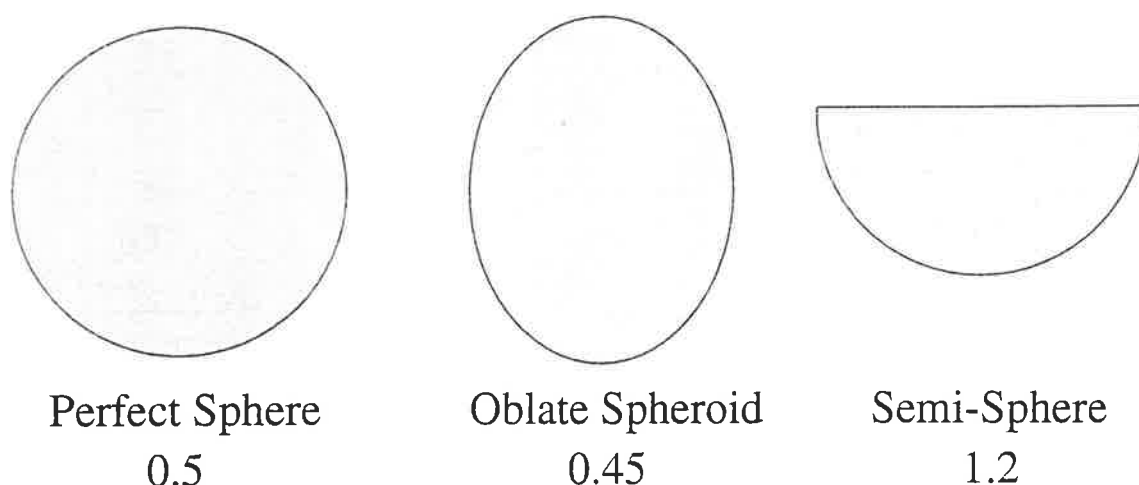


Figure 3.14: Drag coefficients for a few balloon shapes in a flow of Reynolds number $R_e \sim 10^5$, with motion in the direction of the top of the page.

the pressure trace, particularly in the upper troposphere and lower stratosphere...". It is a relatively simple manner to construct a profile of C_D which satisfies nearly every observed ascent rate profile, and it is our contention that variations in drag produce at least the large-scale fluctuations in ascent rate. Whether the small scale ascent rate fluctuations are drag-induced is a more difficult question, which cannot be answered at present. We can say, however, that the magnitude of the fluctuations is not inconsistent with drag-induced causes. The question of a mechanism to produce these variations in C_D in the physical atmosphere has not been fully investigated; indeed, the problems associated with modelling the changes in Reynolds number (and hence C_D) around the critical region may well leave a full solution intractable.

As far as the study of atmospheric vertical velocity is concerned, it is unlikely that perturbations of less than a couple of metres per second could be considered as wind variations, definitely distinct from other flow-related variations. Small-scale gravity waves are almost certainly hidden within the large-scale flow variations. One possible exception is the study of lee waves, which tend to be coherent for a number of wavelengths in the vertical, and possess relatively large amplitudes. Applications to lee wave studies, such as those referenced above, provide evidence of such coherent oscillations which can be compared to other (non-drag influenced) measurement techniques. However, other fluctuations of similar magnitude occur which cannot be linked to orographic sources; nearly every published source notes that only a fraction

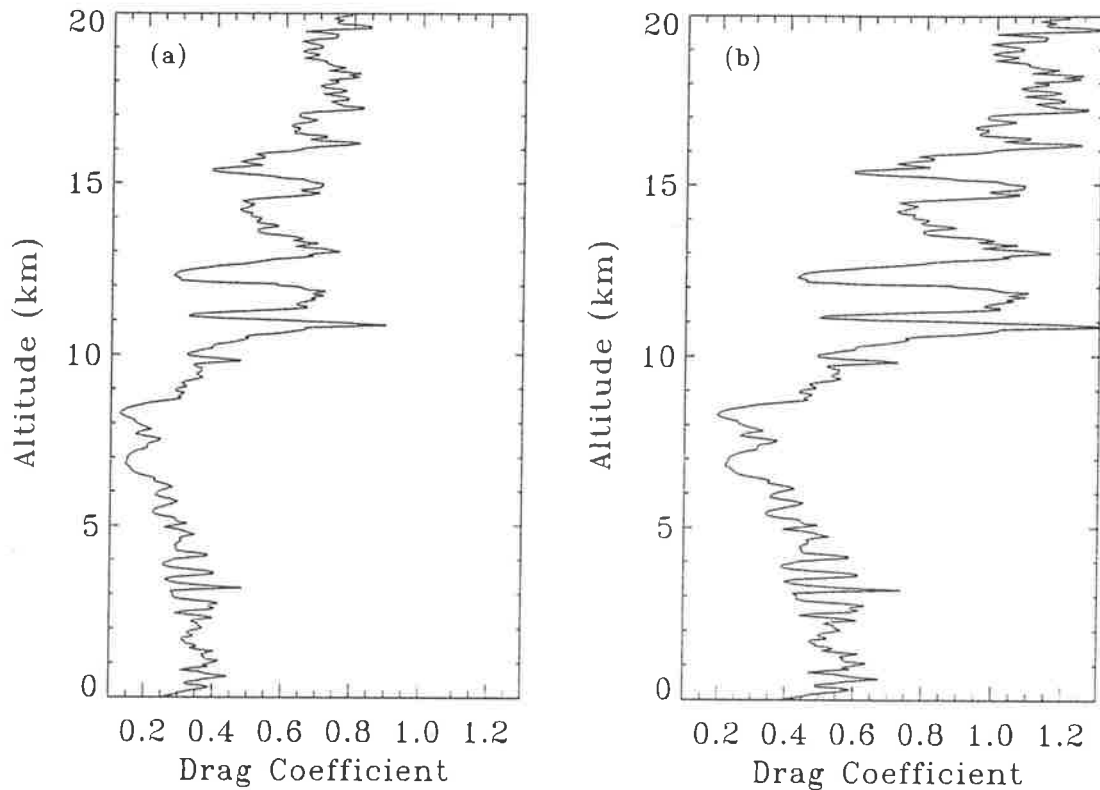


Figure 3.15: Drag coefficient fluctuations required to explain a typical ascent. The initial balloon radius is: (a) 60 cm; and (b) 70 cm.

of the collated ascent rate profiles can be linked to lee waves. Unpublished material from the MU radar in the mountains of Shigaraki, Japan, consistently show similar large-magnitude fluctuations which are only sporadically associated with regular lee wave activity (P. May, personal communication, 1995). An examination of the radar vertical wind profiles acquired around the time of our own balloon ascents show no correlation with the ascent rate variations, regardless of the averaging scheme employed (the profiler vertical velocities are discussed in more detail later); note that orographic effects are minimal in the vicinity of the field site, and that wind variations aloft should be purely non-orographic in nature.

Another possibility that should not be ignored is the feedback effect of drag variations on the atmosphere; strong gravity waves may well produce an environment with regular flow variations, so that C_D and hence observed ascent rate vary in a consistent, predictable manner. The possibilities for feedback are obvious: as ascent rate

increases, the balloon will tend to flatten in the direction of travel, increasing drag and decreasing the ascent rate; a decrease in ascent rate may change the Reynolds number of the flow sufficiently to change C_D significantly, which will similarly alter the ascent rate. These effects generally appear to be of a negative feedback nature, consistent with the observed near-constant mean ascent rates.

There are at least two extensions which can be made to our model, but both are found to be of small effect. Firstly, consider the concept of superpressure. The elastic material of the balloon will exert a force on the interior gas, so that the internal pressure is greater than the external (environmental) pressure by some amount called the superpressure. This concept is dealt with in many materials engineering texts (see, for example, *Gent [1992]*); in our case, the superpressure is given by:

$$p_s = \frac{\mathcal{G}\varepsilon}{R_u} \left[\frac{R_u}{R} - \left(\frac{R_u}{R} \right)^7 \right] \quad (3.24)$$

where $\mathcal{G} = NkT \approx 0.3 \times 10^6$ Pa is the shear modulus, $\varepsilon \approx 0.15 \times 10^{-3}$ m is the balloon material thickness, $R_u \approx 0.285$ m is the unstretched ("just filled") balloon radius, and R is the actual balloon radius (as above).

In practice, if the superpressure is large, the balloon will expand more rapidly than originally expected, reducing the expected ascent rate especially at the upper levels. Estimates of typical superpressures during an ascent from ground level to 20 km altitude vary smoothly between about 1 and 0.2 hPa, or less than 0.5%. This corresponds to an additional increase in radius of between 1 and 5 cm, or about 2%, over the same range. This is significantly less than our error in determining the initial balloon radius (which is around 10%, as explained above). The effects on terminal velocity are negligible. While on the subject of the balloon material, gas leakage through the elastic membrane is a factor which would affect the balloon radius as a function of time (height), although the effects are again likely to be negligible.

Another extension to the model involves temperature lag. There will be a differential between the temperatures inside and outside the balloon, since the ascent rate is relatively rapid. For example, at typical ascent rates of about 5 ms^{-1} the environmental temperature will drop about 10 K in three minutes, whereas the e -folding time for thermal equilibrium (via a simple lumped capacity heat transfer calculation, ignoring

convection) is of the order of sixteen minutes [*Lalas & Einaudi*, 1980]. By this time, of course, the balloon has moved into an even colder (higher) region.

A full analytic solution of the equations of motion taking temperature lag into account is possible, but complicated by feedback effects; investigation of the problem is ongoing. A simple illustrative variation to the model involves a straightforward linear time lag, such that the temperature of the balloon gas as a function of time is given by $T'(t) = T(t - t_0)$, where t_0 is the lag time. In the troposphere, the balloon will generally be warmer than its surroundings, and ascend faster than expected; in the stratosphere, it will be cooler and slower. This alters the buoyancy force acting on the balloon by a factor T'/T , as a simple reformulation of Equation 3.12 making no assumptions regarding states inside and outside the balloon shows. Taking t_0 to be about ten minutes, the balloon will be between 0 and 20 K (up to 8%) warmer than its surroundings in the troposphere, and between 0 and 10 K (up to 5%) cooler in the stratosphere. Applying this to our example, which has a typical troposphere and near-isothermal lower stratosphere produces differences of less than 5% in ascent rates.

The model formulation may be made more rigorous by considering the effects of water vapour (or mixing ratio, q) through the use of virtual temperature:

$$T_v = T \left(\frac{1 + q/0.622}{1 + q} \right) \quad (3.25)$$

Typical mixing ratios are less than 1% (10 g kg^{-1}), producing negligible differences between T and T_v for our purposes. Another problem is the possibility of ice forming on the balloon as it collects water vapour and moves into cooler regions, which would alter the total mass of the system. This has not been examined, and awaits further examination; however, there may well be a self-limiting effect in action, as most balloons which aggregate ice are subject to bursting due to decreased flexibility. With typical freezing levels near 5 km altitude, it is unlikely that a severely iced balloon would reach altitudes of interest.

Carlson & Horn [1983] look at the problems of drag and thermal balance for large balloons (radius 10 to 25 m), and provide a number of references to work in that field. They feel that variation in drag coefficient is the major factor below the

tropopause, while the initial (filling) mass dominates above the tropopause. The degree of applicability of their results to small meteorological balloons is not well understood at present.

To summarise, then, the extraction of neutral air motions from balloon ascent rates is fraught with difficulties. Exceptions to this occur in regimes of strong, persistent oscillations such as lee waves, where the magnitudes (possibly together with feedback effects) serve to regulate the environment to a significant degree. The question of drag coefficient variation is perhaps most deserving of further study; progress on that front has applications across the spectrum of balloon-ascent related fields.

3.5.3 Radar Measurements

Profiler estimates of the vertical velocity, w , require some discussion as they present a unique way to measure this meteorologically interesting variable. The measurement can either be direct (with a vertically-pointing radar beam, so that $V_r = w$) or indirect (derived from VED or VAD observations, for example). Ascertaining the validity of the measurement is difficult, since the standard techniques for estimating w outlined above (Section 3.5.1) involve derivations from large-scale synoptic analyses, which obviously return a broad-scale temporal and spatial average. Comparisons have been made, and show good agreement in general terms [Nastrom, 1984; Nastrom *et al.*, 1994] although there can be contamination from lee-wave effects near mountains [Nastrom *et al.*, 1985a; Larsen *et al.*, 1988] or heavy rain [Balsley *et al.*, 1988]. Nastrom *et al.* [1990] consider the statistical uncertainties in mean vertical velocity due to finite record length and data gaps, and conclude that natural geophysical variability places a limit of about 2 cm s^{-1} on radar-determined average w over an hour or more. On the mesoscale, case studies have shown that radar vertical motions are consistent with established models of the secondary circulation around frontal systems, jets, and the like [Green *et al.*, 1988; Nastrom *et al.*, 1989; Crochet *et al.*, 1990]. Our own comparisons are presented later (Chapter 6, for example), and provide further convincing evidence for the applicability of profiler vertical winds.

Synoptic analyses and long-term profiler averages (as referenced earlier) reveal

mean vertical velocities of the order of 5 cm s^{-1} over a few hours; natural variability, usually associated with high-frequency gravity wave activity, produces a standard deviation of the order of 20 cm s^{-1} over the same timescale (mesoscale events can, of course, generate vertical velocities more than an order of magnitude larger). Horizontal winds on the same spatio-temporal scales average around $10\text{--}20 \text{ m s}^{-1}$, large enough to significantly contaminate vertical velocity estimates if the “vertical” beam is not directed exactly towards the zenith. As discussed earlier, broad beamwidths can also introduce errors due to horizontal velocity components contributing from the extremes of the beam. Radial velocities on vertical beams could also be affected if the aspect sensitivity function of the scatterers was not centered within the beam [Röttger, 1981]. This “tilted layer” effect, which has been discussed in some detail in the literature, has been extensively examined with interferometric techniques [Röttger & Ierkic, 1985; Larsen & Röttger, 1991; Palmer *et al.*, 1991; May, 1993a; Larsen & Röttger, 1993; Palmer *et al.*, 1993]. Vertical beam and VAD estimates of w show some degradation of agreement in regions of anisotropic scatter, although averages over a few hours give consistent results [Larsen *et al.*, 1991; Fukao *et al.*, 1991; Yoe & Rüster, 1992]. Measurements of w by McAfee *et al.* [1994] with 50 MHz and 404 MHz profilers—where the latter, operating in the UHF band, should not be affected by anisotropies such as layer tilt—display good agreement over time. Regions of difference do exist, however, predominantly in regions experiencing hydrometeor contamination (a greater problem at UHF than VHF). Obviously, there remains much work to be done in this area. As a final note on vertical velocity biases, Nastrom & VanZandt [1994] and Hoppe & Fritts [1995] recently compared gravity wave activity, mean vertical velocity and gravity wave activity (as measured by vertical velocity variance) and found a correlation between greater velocity variance, increased reflectivity, and a negative bias in vertical velocity.

It should be pointed out that the biases which are discussed above are of the order of a few centimetres per second at most, when averaged over typical timescales of an hour or more. This is comparable to geophysical limitations (Nastrom *et al.* [1990], as described above), and to the precision of an individual measurement (*e.g.* from Equation 2.11, a power spectrum with n_s spectral points will have a spectral resolution of $|V_r|_{\text{max}}/(n_s/2) = \lambda f_p/2n_s n_c$; for the typical parameters outlined in Table 3.1, this is

about 4 cm s^{-1}), although averaging will improve the precision of estimates. Regardless, the possibility of these biases affecting measurements of w should be borne in mind when interpreting profiler vertical velocities, especially long-term averages.

We note that the spaced antenna method, through the Full Correlation Analysis, can be used to determine vertical as well as horizontal velocities provided we use the entire complex signal and not just the amplitude information. However, the generally wide antenna beamwidths in SA receiving systems are especially prone to horizontal wind contamination as discussed for DBS systems. Also, if the SA system is bistatic (physically separated transmission and reception arrays) rather than truly monostatic (transmission and reception at the same location), leakage of the horizontal wind into the vertical velocity estimates can occur [Vincent *et al.*, 1987; May, 1993a]. Care must be taken in the interpretation of SA/FCA-derived vertical velocities in such cases.

Chapter 4

ESTIMATING PROFILER PERFORMANCE

Anyone who conducts an argument by appealing to authority is not using his intelligence; he is just using his memory.

— *Leonardo da Vinci*

4.1 Introduction

The measurement of winds in the troposphere and lower stratosphere provides fundamental data for use in global circulation and forecasting models. The world network of balloon-borne radiosonde launching sites provides these measurements, using a variety of techniques based on calculating the change in position of the sensor as it ascends through the atmosphere. There are many limitations to this arrangement: in particular, the sparse distribution of launching stations in many parts of the world, and the low frequency of launches. This restricts forecasting efforts on both spatial and temporal scales.

As we have already discussed, VHF wind profilers are capable of measuring winds in the atmosphere from backscatter from the clear air. Most operational systems make use of the Doppler beam swinging (DBS) technique, where the radar beam is tilted about ten degrees off-vertical in a number of azimuths, and the radial velocity at each azimuth is used to infer the horizontal wind field. More recently, the spaced

antenna (SA) technique, where the diffraction pattern on the ground due to moving irregularities in the atmosphere is sampled to determine the mean wind, has been proposed as an alternative. A major advantage of the SA technique is that it makes use of a purely vertical beam which can benefit from enhanced backscatter from stable layers, as outlined earlier. These radars are capable of operating continuously and unattended, sampling a vertical column of the atmosphere with a time resolution of minutes and a height resolution of a few hundred metres. Because of this, profilers are suited to studying meteorological phenomena on the mesoscale as well as the synoptic scale. This is particularly important for short-range forecasting, where the monitoring of events in close to real-time is desirable.

Recognising their utility, increasing interest has been shown in using networks of wind profilers, especially in regions where mesoscale events are frequent. Profiler networks have been evaluated in Colorado [*Strauch et al.*, 1984], France [*Pilon*, 1993], and across the Pacific [*Gage et al.*, 1990]. However, before profiler winds can be accepted as a replacement for selected rawinsonde stations, rather than simply an adjunct, they must satisfy two criteria: accuracy of the retrieved winds, and continuity in time and height of the returns.

In contrast with the extensive treatment of profiler wind accuracy in the literature reviewed earlier, examinations of profiler altitude coverage (and the associated variation with time) are quite limited. *Frisch et al.* [1986] describe the behaviour of the Colorado network on statistical basis and, in later work, consider seasonal and annual variations [*Frisch et al.*, 1990]. *Nastrom et al.* [1985b] calculate the average daily returned power at four locations, and associate the variations with gravity wave activity and afternoon convection. Sporadic reductions in vertical beam returned power (“biteouts”), generally just below the tropopause, have been observed by a number of authors (*e.g.* Figs. 1 and 4 in *Eckermann & Vincent* [1993]: Fig. 1 in *May* [1993b]), and are generally ascribed to the lack of moisture, near-adiabatic temperature lapse rates, and decreased turbulence at upper levels. These biteouts are a major concern in the design of operational profilers, and we shall attempt to quantify their occurrence. Operational usage requirements generally state that profilers must be capable of returning data through to the tropopause level continuously at a given averaging

interval (usually an hour or so). System design should take this into account, and we shall show how local conditions can affect performance in a predictable manner.

4.2 Methodology

As part of the South Australian Fronts Experiment (SAFEx), a wide range of atmospheric conditions were sampled with a 54.1 MHz VHF wind profiling radar and co-located MARWIN radiosondes. Nearly 150 wind- and thermodynamics-measuring sondes were launched from the Buckland Park radar site, at approximately three-hour intervals during seventeen separate events; flight details are listed in Appendix C. A typical balloon takes about 90 minutes to reach 25 km, which sets an upper limit on the rate of launches since only one sonde can be tracked at a time. An ideal Intensive Observation Period (IOP) during SAFEx would be of 36 hours duration, with flights every three to four hours. It should be noted that the SAFEx launches were usually made into conditions some eighteen hours either side of frontal passages, with a relatively rapidly evolving atmosphere. Hence, short-term variations can be observed and compared in detail. By way of comparison, launches from the Bureau-operated sites (for which we make our predictions) are made either 12- or 24-hourly.

During most of the observational programme the profiler was operated on a five minute cycle, alternating one vertical Doppler retrieval (64 seconds of data collection, plus analysis time) with one spaced antenna run (128 seconds plus analysis). The results presented here use returned power ($\mathcal{P}$) and signal-to-noise ratio (SNR, or $\mathcal{S}$) from the vertical Doppler beam in this observation mode alone: that is, 64 seconds of data every five minutes. Hence, our base predictions are valid for this temporal resolution; the effect of different sampling is treated later.

Our approach is as follows: estimates of the gradient of generalised radio refractive index from the high-resolution thermodynamic sonde data allow for an effective calibration of the radar system by comparison with returned power profiles. Based on this calibration, and a derived returned power/signal-to-noise ratio, it is possible to make predictions of the performance of wind profilers at any location where routine historical radiosonde data or climatological thermodynamic information are available. We

examine a number of sites, and compare our predictions for Darwin with actual profiler observations there. In terms of operational profiler requirements, the characteristics of the region just below the tropopause present the limiting case.

The range of conditions which were observed represent a wide cross-section of the expected annual variation, and allows us to extend our results to a variety of locations and climates.

4.3 Calibration

Our aim here is to find a quantitative relationship between the SNR, $\mathcal{S}(r)$, attained with the vertical radar beam, and the sonde (p, T, q) data. To this end, we note the proportionality between returned power and M^2 expressed by Equations 2.7 and 2.21, and that between $\mathcal{S}$ and returned power:

$$\mathcal{P} = \mathcal{S} \times k_B B (T_c + T_{rx}/\alpha)/d \quad (4.1)$$

where k_B , B , and d are the Boltzmann constant, spectral bandwidth, and duty cycle respectively, T_c and T_{rx} are the cosmic and receiver noise temperatures, and α is the radar antenna/transmission-line system efficiency. This pair of relationships can then be inverted to estimate $\mathcal{S}$ from a climatology of the thermodynamic variables. Note that Equation 4.1 implies that the proportionality is constant with height—it disregards range-dependent effects (such as clutter), which are assumed to have been removed from the $\mathcal{P}$ and $\mathcal{S}$ estimates. Since clutter contamination is not removed by the Buckland Park signal processing software, some degree of care is required.

In any comparison between fundamentally different sensors, it is important to ensure that the measurements are in fact comparable. The profiler vertical beam samples a purely vertical profile of the atmosphere every few minutes (*i.e.* it is an Eulerian sensor), and the returns are a volume-average from within each range gate. Balloon-borne radiosondes travel with the mean flow, and make a series of essentially point measurements along their path (a Lagrangian measurement). This introduces a number of complications, in particular the fact that the two systems are sampling different parts of the atmosphere in a horizontal plane at any given altitude. For

example, local conditions see our balloons ascend at a finite rate into a wind field that is predominantly eastward (usually with a closed tropopausal jet); hence, the horizontal difference in position increases with the altitude of the measurement. Depending on the strength of the tropopausal jet, the maximum separation of the sampling volumes at an altitude of 20 km can range from tens to hundreds of kilometres; jets approaching 100 ms^{-1} (360 km hr^{-1}) at 10 km altitude have been observed on occasion.

All derivatives calculated from the sonde data were evaluated from the 50 m resolution data using a three-point Lagrangian interpolation method, from which profiles of M^2 (at 50 m resolution) were calculated. To match with the radar sampling resolution a 500 m running mean was applied, and the resultant $\overline{M^2}$ profile was resampled at 500 m intervals, centred on the radar range gates. The Fresnel scatter model predicts that returns from a range gate are predominantly due to partial reflection from gradients in refractive index (high values of M^2) distributed within the range gate volume; turbulent scatter models are similar, in that returns are due to volume-averaged turbulent irregularities. Our averaging and sampling method bears this in mind. Averages were carried out on the logarithmic (dB) values, since the distributions of returned power and M^2 are log-normal [Nastrom *et al.*, 1986], and statistical analysis is usually best performed on normally-distributed data. In addition, any proportionalities are then manifested as linear shifts.

The questions of spatial and temporal separation require more consideration. We have discussed the subject in Section 3.4, where the concept of a “sliding” average was introduced: essentially, a more representative radar parameter average at a given altitude (given temporal evolution of the atmosphere) is one centred on the actual time that the balloon reached that height. For the case of radar power, the sliding average would be formulated by:

$$\overline{\mathcal{P}(z)} = (1/m) \sum_{t=t'-30}^{t'+30} \mathcal{P}(z, t) \quad (4.2)$$

(altitude in metres, time in minutes, and m radar estimates per hour; returned power in arbitrary units, labelled as dB after taking $10\log_{10}$). We have examined both the centered and sliding methods for this case, and found that the discrepancy between them amounts to a variance of less than 0.5 dB in predicted SNR; for consistency, we

use the sliding time-average exclusively.

4.4 Results

Some example comparisons between returned power $\mathcal{P}$ and $\overline{M^2}$ are displayed in Figure 4.1, where the power has been range-corrected (multiplied by r^2), converted to log-scaled units, and matched to the corresponding $\overline{M^2}$ profile in the upper troposphere via a least-squares method. The (altitude-constant) shift which minimised the difference between the two profiles in the upper troposphere, over a 5 km band beneath the tropopause, is indicated. We expect the scattering to be consistent in this region, since it is climatologically dry and dominated by non-specular returns. Note that the x-axis is in $\log_{10}$ units for clarity; multiply by 10 to get dB.

There are three main types of comparison, differing in their behaviour in the stratosphere. One shows a good match through the troposphere and stratosphere (*e.g.* Figure 4.1(1),(2),(4)), while the second indicates that the returned power is lower above the tropopause than would be expected from the M^2 data (*e.g.* Figure 4.1(9),(12),(14)). In approximately half of the latter cases, there is a return to a good match higher in the stratosphere (*e.g.* Figure 4.1(10),(11),(16)); this is our third case. The dry part of $\overline{M^2}$ (M_d , from Equation 2.18) is also shown, to indicate the moisture contribution. For reference, of the 53 flights in SAFEx Phase IV (August to October 1992), 27 flights were classified as “good”, while fourteen of the other 26 flights displayed the “recovery” characteristics of the third case.

We ascribe these variations to changes in the nature of the scatterers between the troposphere and stratosphere. Note that we have assumed an r^2 range-dependence of returned power, which is typical for volume scatter effects. Point targets, however, would have an r^4 dependence, and there is a continuum of possible exponent values in between. If we apply a separate range-dependence in the stratosphere to the cases similar to Figure 4.1(9), then the exponent of r that best fits the data is approximately 2.4. Simulations by *Rastogi & Scheuchner* [1990] indicate that an exponent of 2.22 is appropriate for a planar layer of scatterers (not taking reflection processes into account); the predominance of anisotropic or specular layers in the stratosphere has

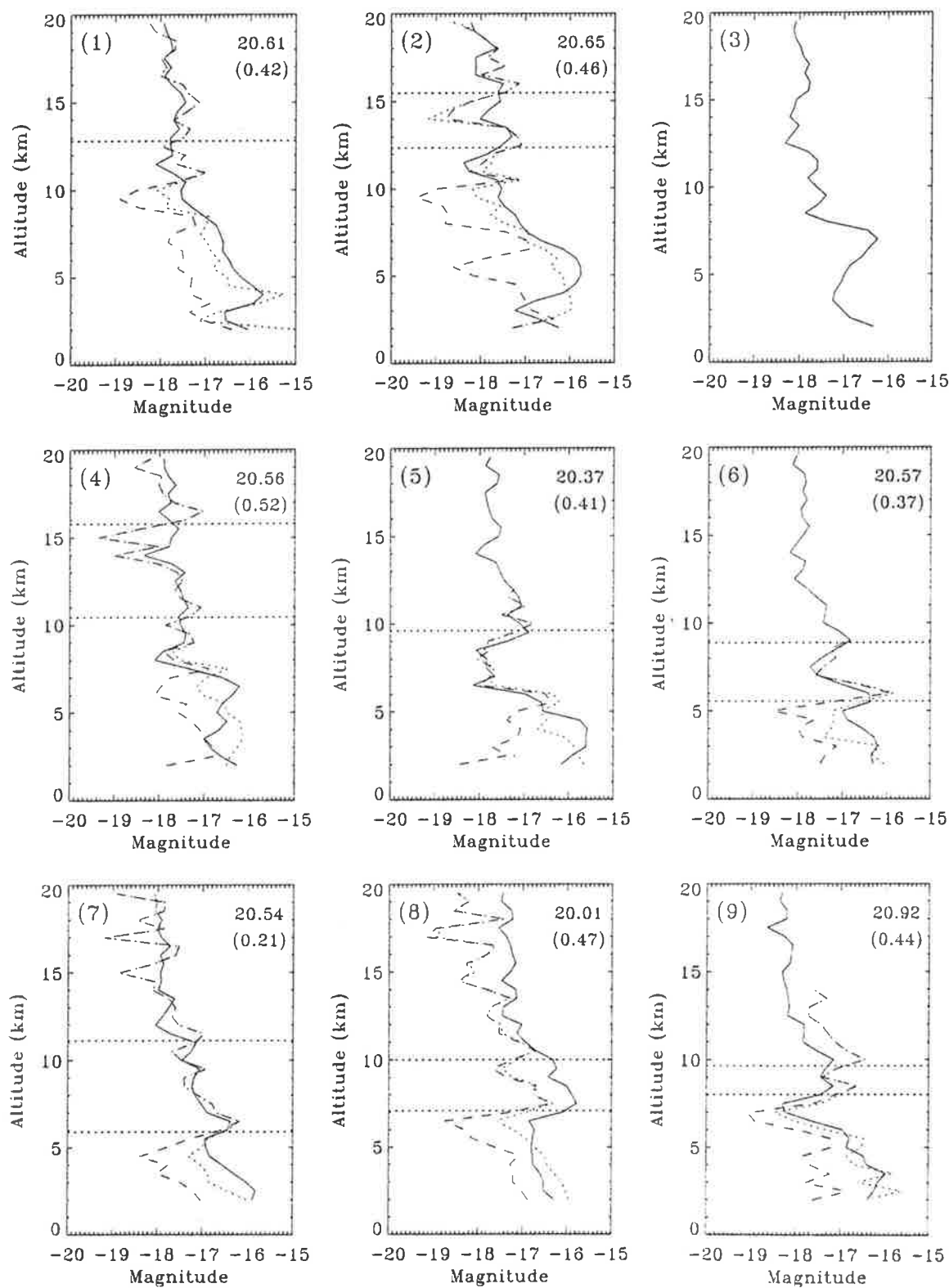


Figure 4.1: (1)–(18): A comparison of returned power and M^2 for the eighteen flights of SAFEx Phase IV, IOP 3. Solid, dashed and dotted lines are, respectively, range-corrected returned power ($r^2\mathcal{P}$), the dry component of M^2 (M_d), and the full (wet) M^2 profile. [Caption continued next page]

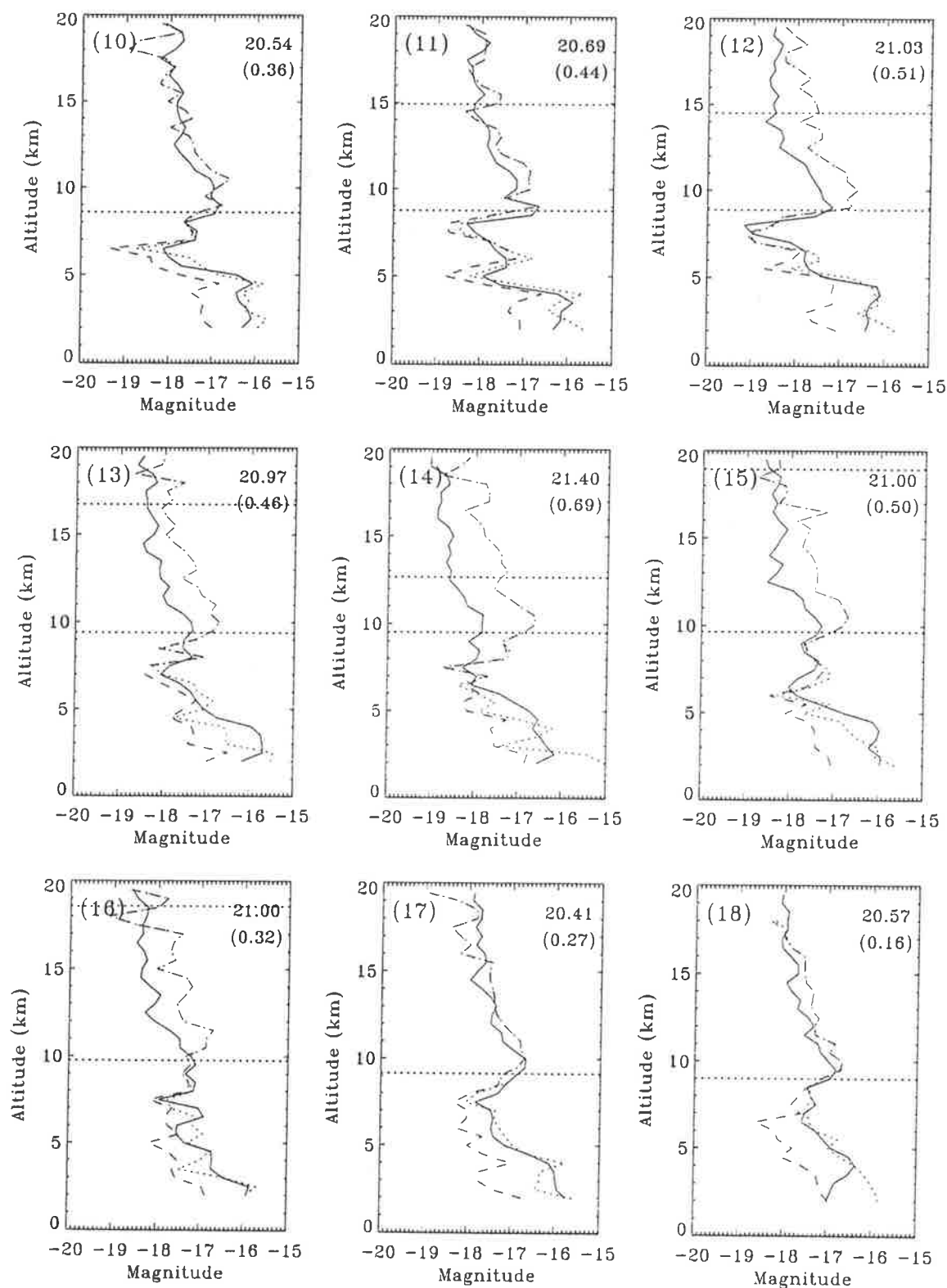


Figure 4.1: (continued) The offset shift applied to the returned power is shown in the top right of each plot (together with the standard deviation in the shift, in parentheses); units are dB/10 to match the plotted profiles. Dotted horizontal lines indicate sonde-determined tropopause heights.

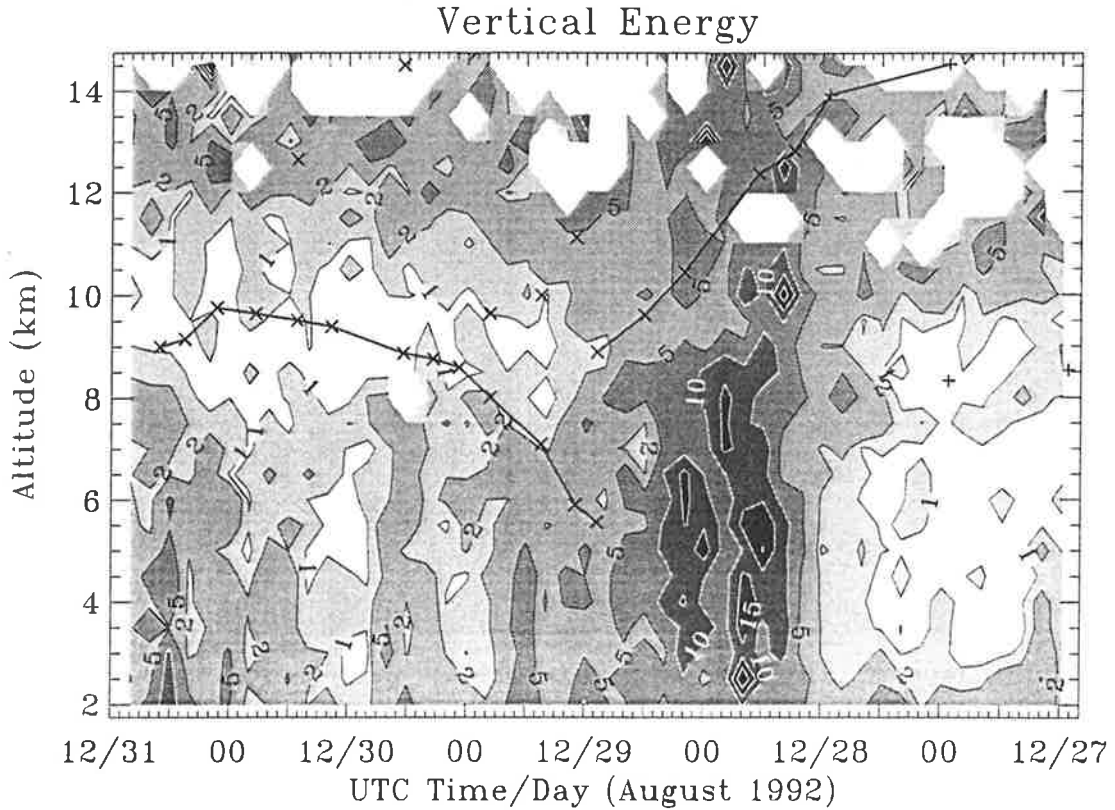


Figure 4.2: IOP IV/3 vertical velocity variance, or energy per unit mass, $\langle w'^2 \rangle$, presented as a two-hour average. Contours are non-linear, with units $\times 10^{-2} \text{m}^2 \text{s}^{-2}$. Note that time runs from right to left as a visualisation aid. Sonde-determined tropopause heights are marked by $\times$ (Buckland Park) and $+$ (Adelaide Airport); a solid line indicates the main tropopause levels around the fold.

been well-documented [Hocking *et al.*, 1986]. In fact, an extension of the latter work [Hocking *et al.*, 1990] demonstrated that scatter became more isotropic above about 18 km, which matches well with our third-case (“recovery”) observations. That is, when the lower stratosphere is dominated by aspect sensitive scatterers, we expect the r^2 -scaled model to underestimate the expected returns from the M^2 model; conversely, when isotropic scatter is dominant, there should be a good match over all observed altitudes.

The third IOP of SAFEx Phase IV (27th to 31st August 1992; hereinafter IOP IV/3) is worthy of note, as it provides a continuous dataset through a variety of conditions. This particular event is discussed in some detail in Chapter 6, but we shall highlight some of the relevant observations here. The initial conditions for IOP IV/3 feature

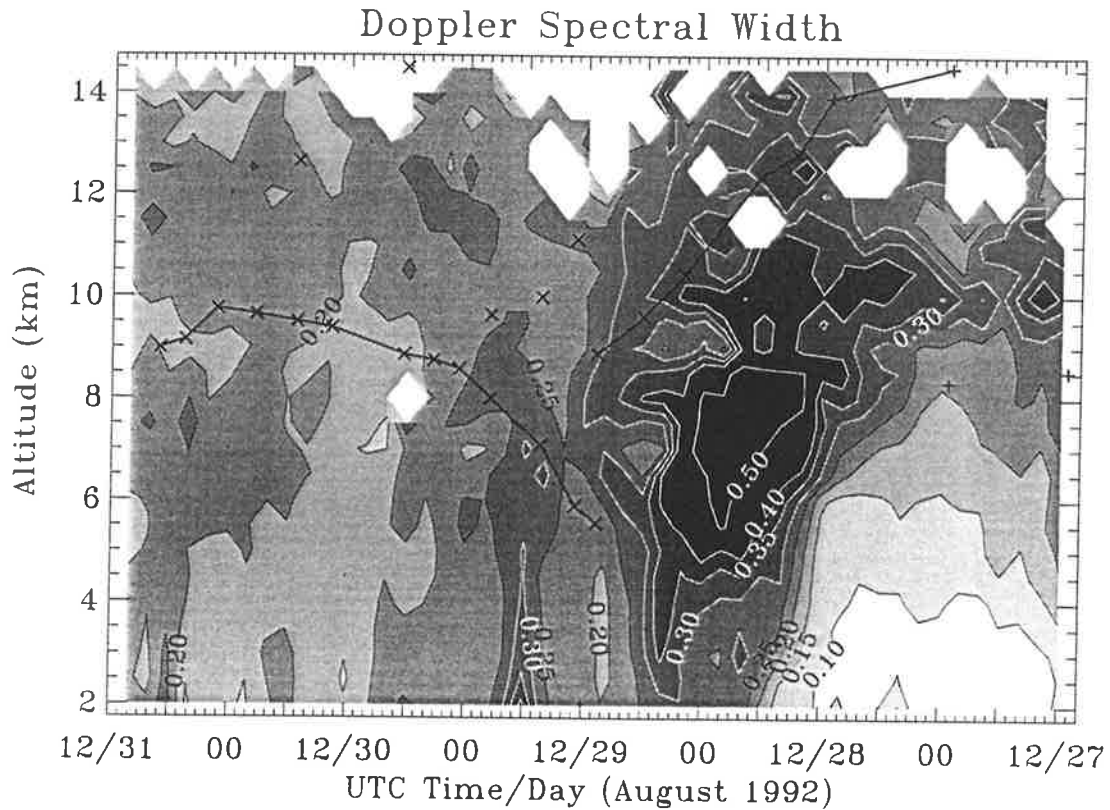


Figure 4.3: IOP IV/3 vertical Doppler beam spectral widths (units of Hz, two-hour average, contour intervals of 0.05 Hz from 0.10 Hz to 0.40 Hz). Beam-broadening and turbulent contributions are included; tropopause markers as in Figure 4.2.

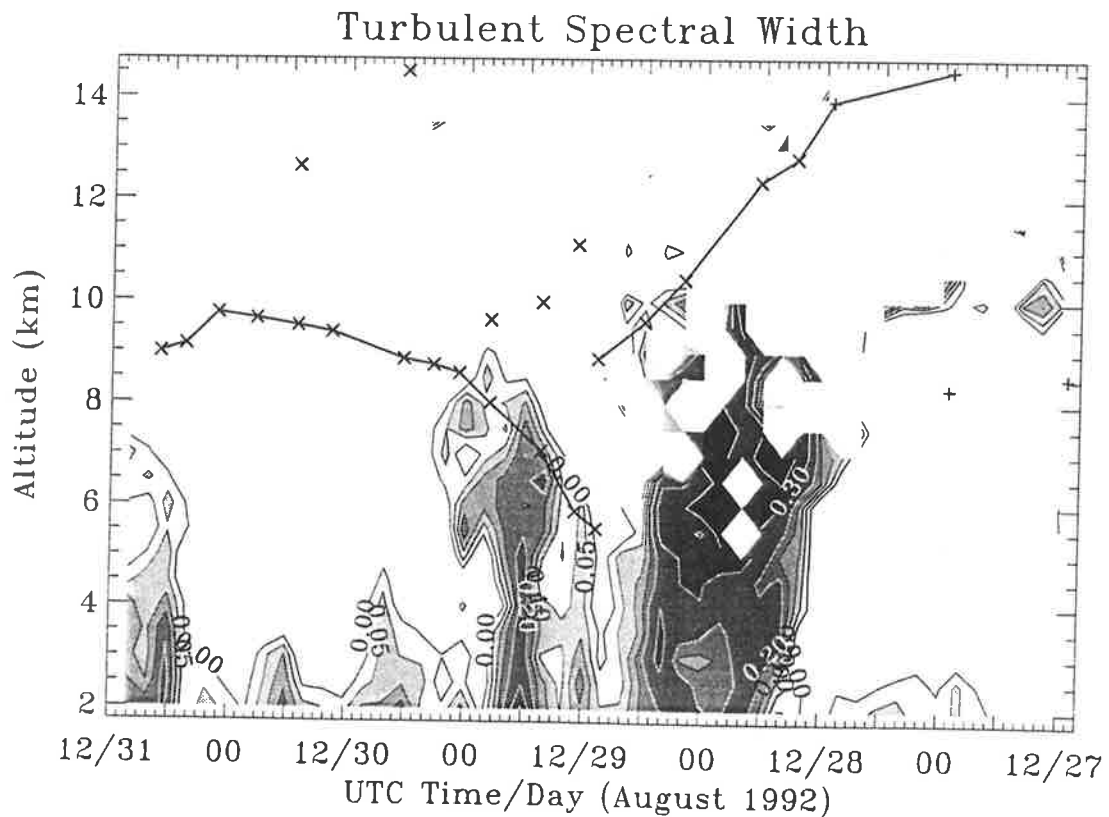


Figure 4.4: IOP IV/3 vertical Doppler beam turbulent contribution to spectral width (units of Hz, two-hour average, contour intervals of 0.05 Hz from 0 Hz to 0.20 Hz). Tropopause markers as in Figure 4.2.

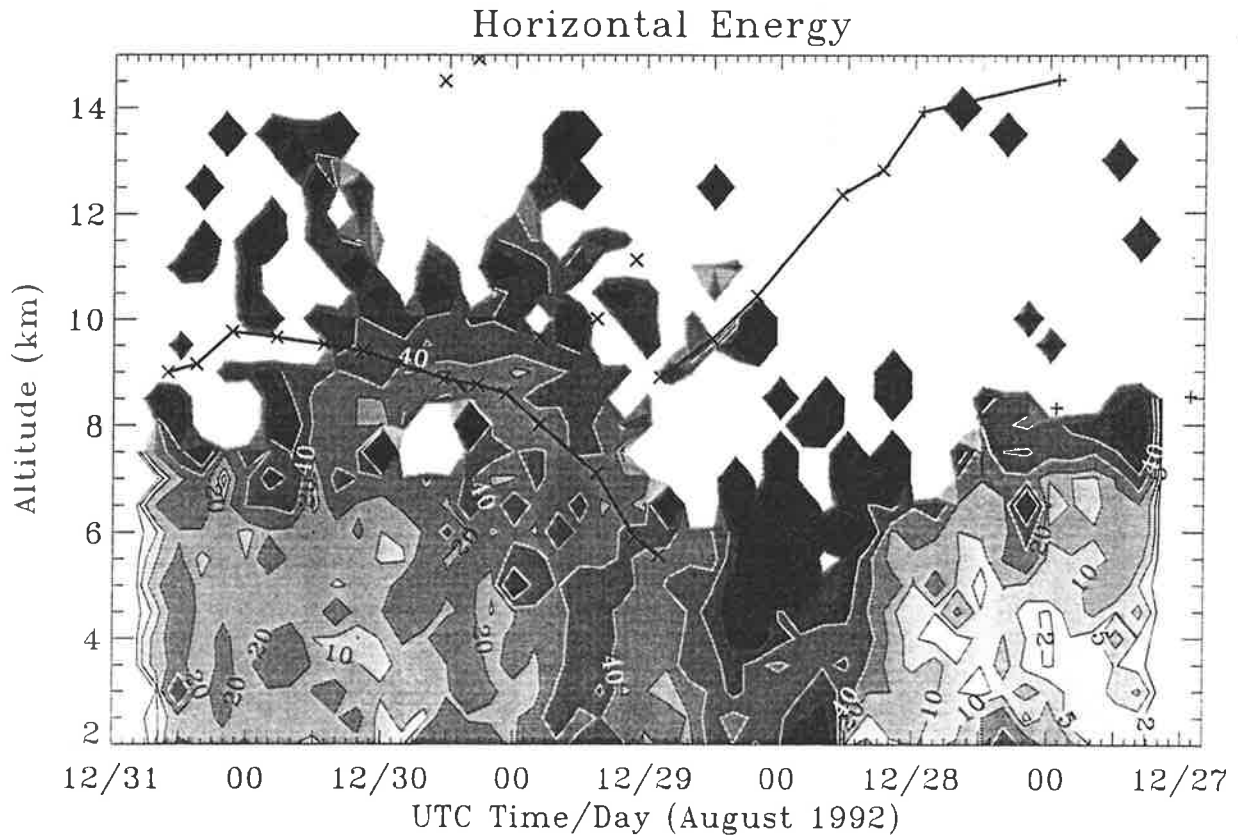


Figure 4.5: IOP IV/3 horizontal velocity variance, or energy per unit mass, ($\langle u'^2 \rangle + \langle v'^2 \rangle$), presented as a two-hour average; periods greater than eight hours have been excluded. Contours are non-linear, with units of m^2s^{-2} . Note that time runs from right to left as a visualisation aid. Tropopause markers as in Figure 4.2.

a strong tropopausal jet, with which one would associate a high degree of turbulence and gravity wave generation (due to geostrophic adjustment, for example). Figure 4.2 shows a high degree of vertical velocity variance in the troposphere (and extending into the stratosphere) around the tropopause fold exit region between 1200 UTC August 28 and 0600 UTC August 29; note in particular the non-linear contours, and the typical “quiet” conditions prevalent some 24 hours earlier. The first five MARWIN flights occur around the fold exit region, and the profiles displayed in Figure 4.1 are typical “good match” cases. The Doppler spectral widths (shown in Figure 4.3) are large in this region and, after removal of the beam-broadened component, retain a significant turbulent component (Figure 4.4). The spaced antenna horizontal velocity variance (Figure 4.5) also reveals enhanced activity around this time; in general, the comparison between turbulent width on the vertical Doppler beam and horizontal

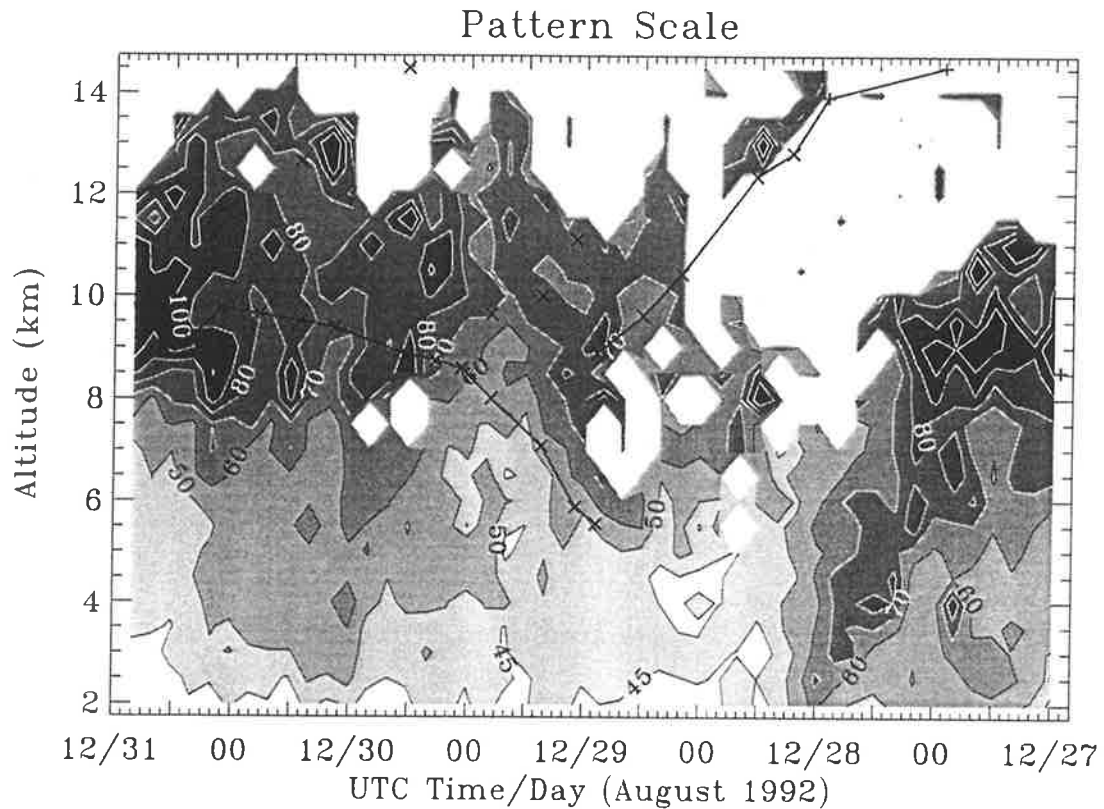


Figure 4.6: IOP IV/3 spaced antenna pattern scale (units of metres, two-hour average, contour intervals of 10 m from 50 m to 100 m). Tropopause markers as in Figure 4.2.

velocity fluctuations is good, at least in the troposphere.

Later, during the passage of the surface event and the tropopause fold (an 18-hour period beginning around 0600 UTC August 29), the vertical velocity variance is far less enhanced, especially in the lower stratosphere where beam-broadening effects dominate the Doppler spectral widths; MARWIN flights 6–9 are examples of the “poor stratospheric match” case, as seen in Figure 4.1. Finally, the last 36-hour period from 0000 UTC August 30 is characterised by a distinct lack of vertical velocity activity near the tropopause, together with narrower spectral widths (longer fading times) and larger spaced antenna pattern scales (Figure 4.6) in the upper troposphere and lower stratosphere. This is indicative of enhanced stability and anisotropic scatter, but a lack of data above 15 km makes it difficult to make conclusive statements regarding the “recovery” region of (presumably) more isotropic scatter in the upper stratosphere. However, the last nine MARWIN flights are generally indicative of “recovery”-type

profiles, strongly suggesting a change in the nature of the scatterers above 15 km or so. It is unfortunate that the spaced antenna records do not extend into the stratosphere to any great extent, since the distribution of horizontal velocity variance would be a useful indication of wave activity in that region.

Due to the variability of cases, we have continued to assume an r^2 range-dependence at all heights: this means that our predictions of SNR in the lower stratosphere may on occasion be overestimated (refer to Equation 4.5). However, since the main region of interest for operational profilers involved in forecasting is up to the tropopause, this is not a serious problem. A possible improvement to the predictions of height coverage could be to determine the mean scatterer nature as a profile of r -exponent varying with altitude, via (for example) a long-term average looking at each month over a number of years. We would expect the average range-dependence exponent to smoothly vary somewhere between 2.0 and 2.22 (although exponents as high as 2.4 would not be unreasonable, given the stratospheric “best fit” mentioned earlier in this section), and could easily be incorporated into the model. The possibilities for such a study could, of course, be extended to an investigation of the nature of the scatterers themselves (see, for example, *Lesicar & Hocking* [1992] and *Allen & Vincent* [1995]). Recent moves by international weather organisations to archive high-resolution sonde data (rather than only the significant levels) would be of major assistance to such a study, providing an excellent basis for comparisons with radar data.

From a total of 53 sonde flights (restricting ourselves to SAFEx Phase IV, August 1992 for consistency, due to radar system improvements over the two-year campaign) the mean shift obtained is 207.0 dB with a standard error of 0.4 dB; that is:

$$10 \log_{10}(\mathcal{P}_r z^2) = 10 \log_{10}(\overline{M^2}) + (207.0 \pm 0.4) \quad (4.3)$$

where $\mathcal{P}_r$ and $\overline{M^2}$ are the measured values (on an arbitrary scale for power), converted via the $10 \log_{10}$ operator to log-scale dB; we use z rather than r for the altitude coordinate to emphasise our vertical-beam assumptions. A selection of IOP-averaged results, where the average $\overline{M^2}$ and $\mathcal{P}$ for the entire observing period are compared, is shown in Figure 4.7. In general, the matches are excellent throughout the atmosphere, with the overall trends and locations of maxima and minima comparing well between

the $\overline{M^2}$ and returned power profiles. This suggests that the extraction of a profile of “scatterer nature” via a point-to-point fit of the best r - (or z -) exponent is feasible; investigation into such a technique is continuing. A variation on this technique is described in Section 5.8.

The power/SNR calibration is performed in a similar manner, by a simple comparison of the log quantities. Note that since we define the signal-to-noise ratio as the difference (in dB) between the signal power and the noise, then the calculated value $\mathcal{P} - \mathcal{S}$ is equal to the noise level; given isotropic sky noise, this should be a constant. The expected linear relationship between $\mathcal{P}$ and $\mathcal{S}$ breaks down because of digitisation effects when the signal is strong (an upper limit of approximately 20 dB is imposed on the recorded SNR), and non-white noise effects (ground clutter, and other range-dependent contamination); but again, in the main region of interest (above 5 km or so), a strong proportionality is evident in Figure 4.8. Using SAFEx Phase IV data for consistency with the derivation of Equation 4.3, the overall result is:

$$10 \log_{10}(\mathcal{P}_r) = 10 \log_{10}(\mathcal{S}) + (9.81 \pm 0.09) \quad (4.4)$$

where both sides of the equation are in dB.

Combining Equations 4.3 and 4.4 allows us to make predictions of SNR from $\overline{M^2}$ profiles, for a radar system identical to the Buckland Park VHF array:

$$\mathcal{S}(z) = 10 \log_{10} \left(\frac{\overline{M^2}(z)}{z^2} \right) + (197.2 \pm 0.5) \quad (4.5)$$

where $\mathcal{S}(z)$ is estimated in dB; again, we denote the altitude coordinate by z . This is, of course, simply an “applied” calibration of the radar system. A fully calibrated system (where returned power is known in absolute terms, and an accurate value of antenna efficiency is available; *e.g.* Eaton *et al.* [1988]) would allow for a general solution, as well as a quantitative comparison with the scattering models described by Equations 2.6 and 2.7. Since receiver gains are adjusted manually on the Buckland Park VHF system, such a full calibration is currently infeasible for the purpose of comparison. However, upgrades in progress on the Buckland Park system will permit calibration, at which point the constants in Equation 4.5 can be confirmed.

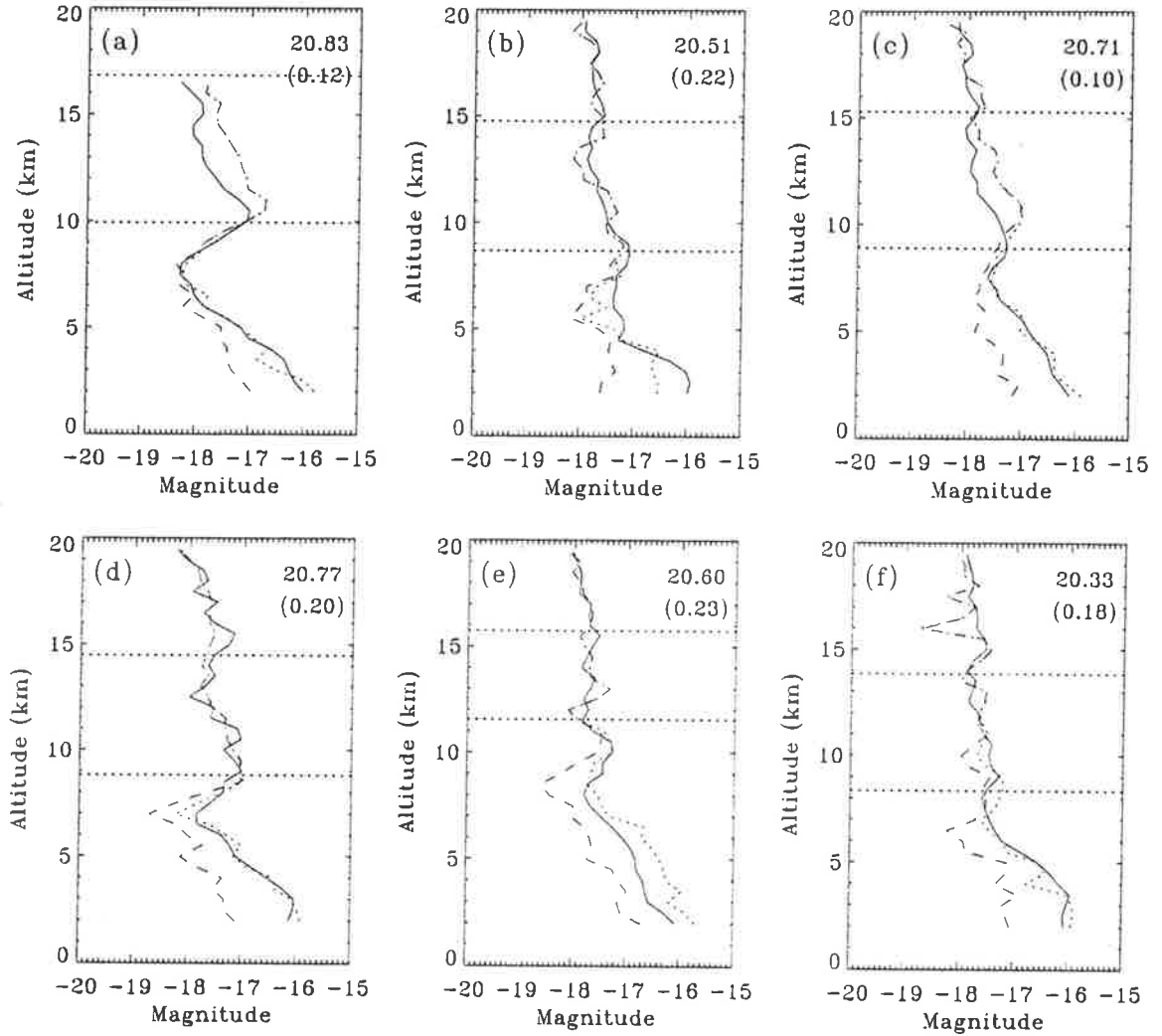


Figure 4.7: A comparison of average returned power and $\overline{M^2}$ for six IOPs from SAFEx Phase IV: (a) IOP 1; (b) IOP 2; (c) IOP 3/4; (d) IOP 5; (e) IOP 8; (f) IOP 11. Details of IOPs are given in Appendix C. Solid, dashed and dotted lines are (respectively) range-corrected returned power ($r^2\mathcal{P}$), the dry component of M^2 (M_d), and the full (wet) M^2 profile. The offset shift applied to the returned power is shown in the top right of each plot (together with the standard deviation in the shift, in parentheses); units are dB/10. Dotted horizontal lines indicate average sonde-determined tropopause heights, determined from a two-cluster grouping.

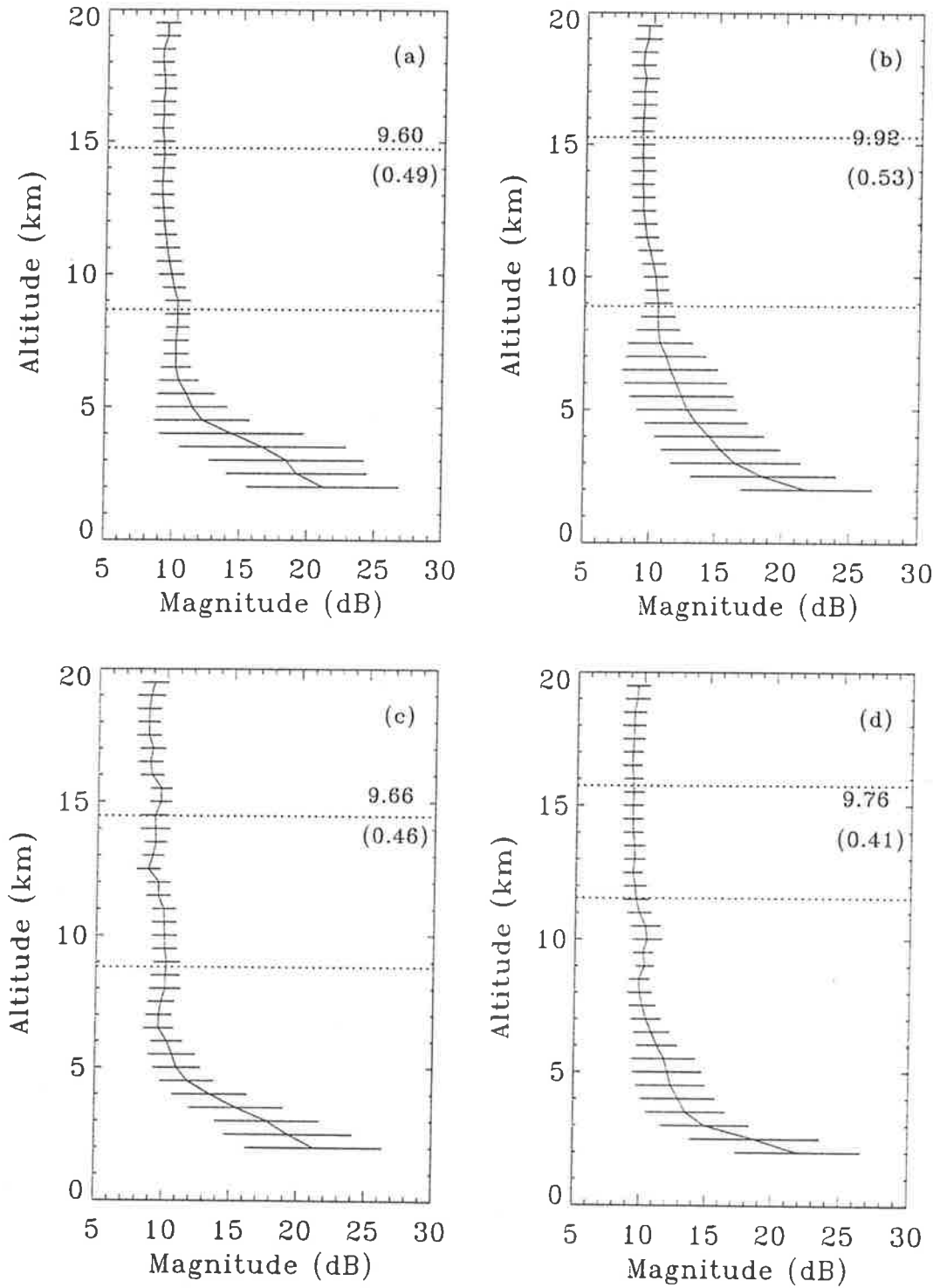


Figure 4.8: A comparison of average returned power and signal-to-noise ratio for some cases from SAFEx Phase 4: (a) IOP 2; (b) IOP 3/4; (c) IOP 5; (d) IOP 8. The profiles show the noise level (*i.e.* $\mathcal{N} = \mathcal{P} - \mathcal{S}$); units are dB (*i.e.* $10\log_{10}$), and the error bars are two standard deviations wide ($\pm\sigma$). Dotted horizontal lines indicate average sonde-determined tropopause heights, determined from a two-cluster grouping. The mean shift (and standard deviation) between 8km and 14km altitude is shown at the top right of each plot.

4.5 Predictions

4.5.1 Adelaide (35°S, 138°E)

The predictions for Adelaide are based on the twelve-hourly Bureau radiosonde flights from Adelaide Airport (AAP), 40 km south of the radar site. Direct comparison with nearly simultaneous sonde flights from the site show negligible differences in the (p, T, q) profiles, as expected from the relatively small separation. This means that the calibration could be performed just as well with the AAP data; similarly, profiler sites can be evaluated using the sonde data from nearby launches. However, we are now concerned with average performance rather than exact matching, so geophysical variation becomes important.

The mean profile of $\overline{M^2}$ for a number of months is shown in Figure 4.9, where the two-cluster average sonde-determined tropopause heights have also been indicated. Although WMO guidelines recommend sonde launches at 0000 UTC and 1200 UTC, local practise is to launch at 2300 UTC (0830 LST) and 1100 UTC (2030 LST). However, during the daylight savings period (mid-October to mid-March), launches are made at the same *local* time; that is, 2200 UTC and 1000 UTC. For clarity, we shall continue to refer to launches by their non-daylight savings UTC times, although comparisons have been made at the correct (local) times. Since we are dealing with processes that may well be affected by local conditions near sunrise, this latter point is particularly important! Note that the 1100 UTC and 2300 UTC sonde flights have been combined here, there being no significant difference between them above the boundary layer; also, the units are dB/10 for clarity. The broad features are common to any month in the dataset: $\overline{M^2}$ decreases close to linearly with height at about 4 dB km⁻¹, has a local maximum at the tropopause with a 10 dB enhancement, then falls off more slowly (but still linearly) in the stratosphere at about 1.5 dB km⁻¹. These trends and values compare well with the observations of Röttger [1980*b*] and Nastrom *et al.* [1986]. The tropospheric variability is about twice that seen in the stratosphere, dropping from about 2 dB to about 1 dB across the (upper) tropopause (expressing the variation as a standard error in the mean $\overline{M^2}$).

The predictions of SNR from an extended dataset of high-resolution sonde flights

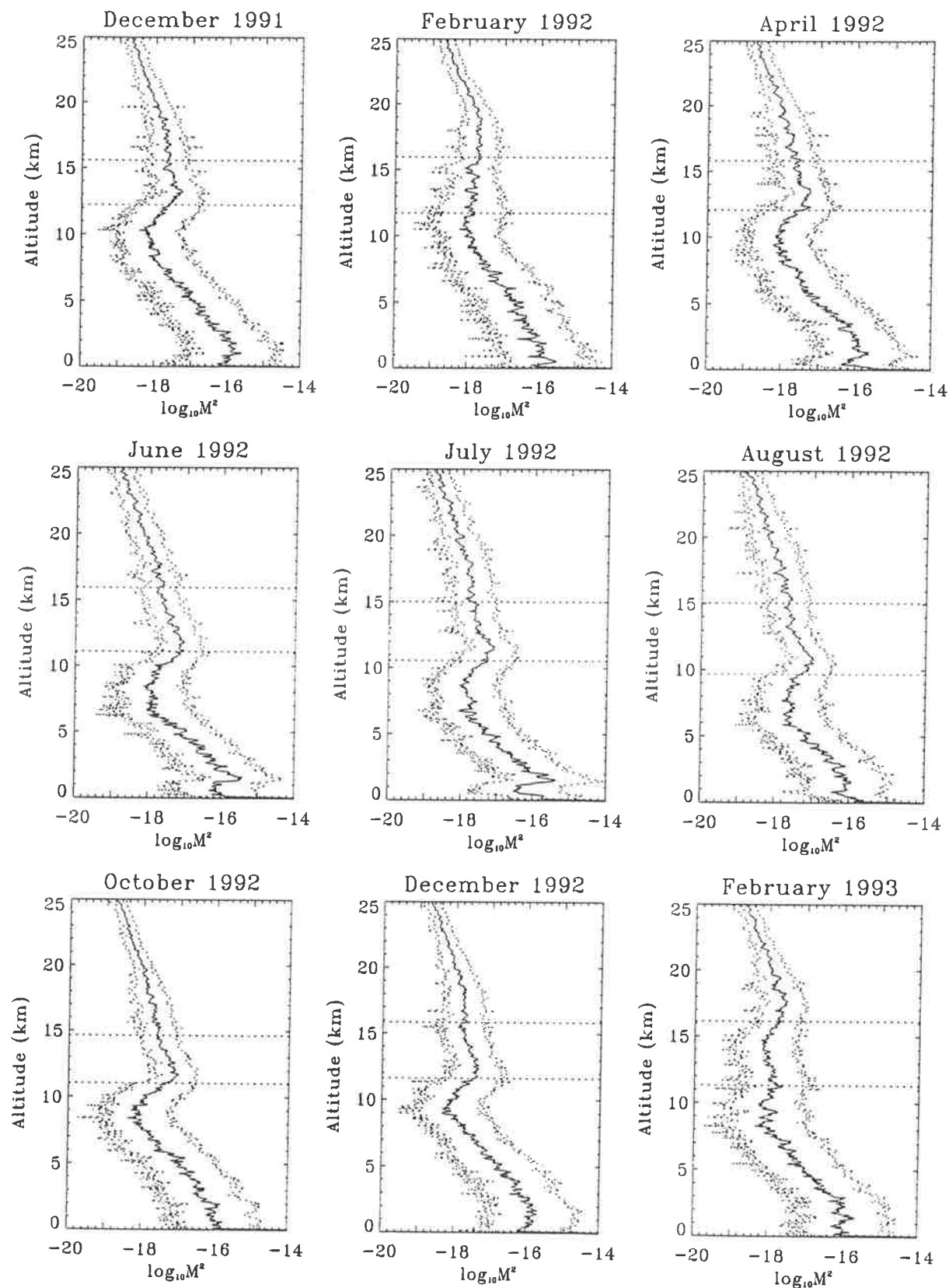


Figure 4.9: Mean profiles of $\overline{M^2}$ from Adelaide Airport sonde data for a number of months. The solid line indicates the mean, while the bracketing profiles (dotted) indicate one standard deviation either side. Dotted horizontal lines indicate average sonde-determined tropopause heights, determined from a two-cluster grouping. Approximately sixty sonde flights contribute to each plot (two flights per day, at 0830 LST and 2030 LST).

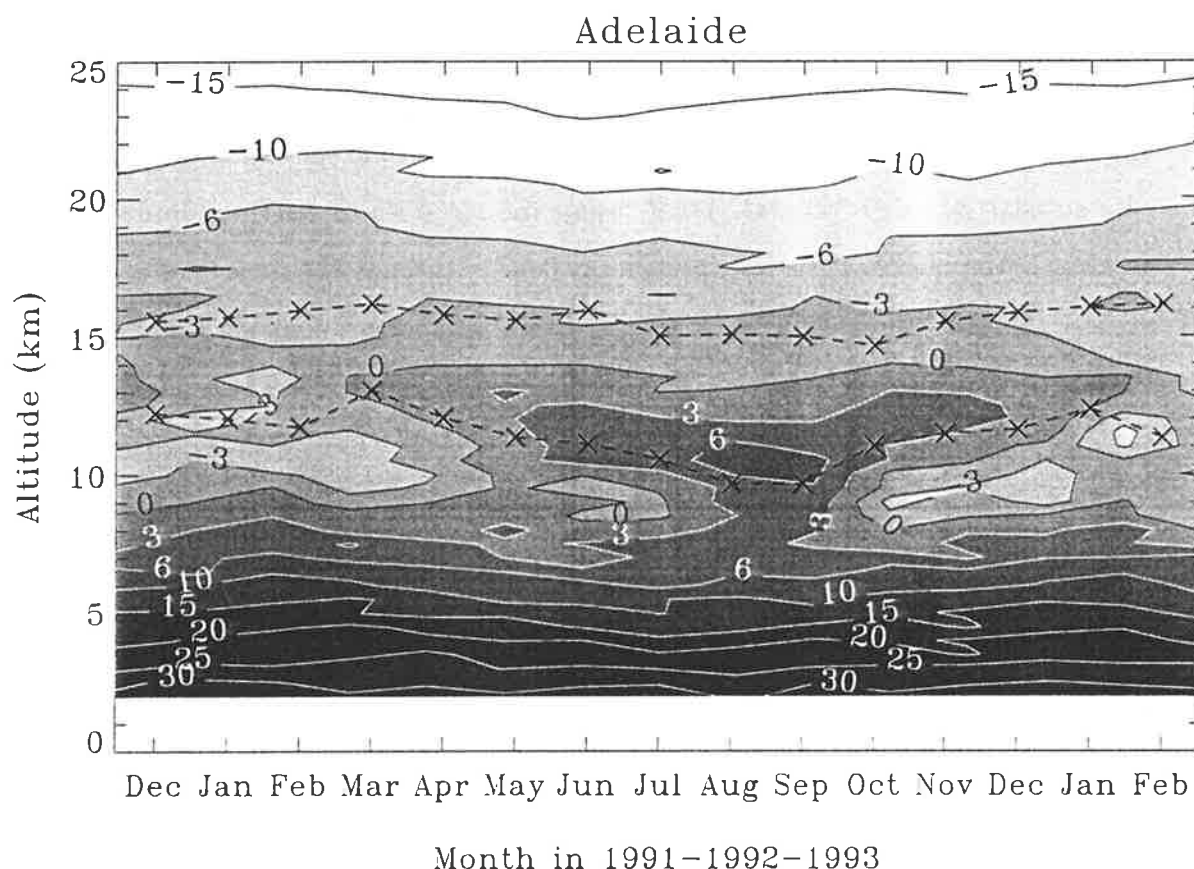


Figure 4.10: Predictions of observed SNR, derived from Adelaide Airport sonde data. Note that contours of SNR (in dB) are non-linear, as an aid to predictive applications. Dashed lines and $\times$'s indicate cluster-mean derived tropopause levels.

from Adelaide Airport are combined in Figure 4.10 as a shaded contour plot. Monthly cluster-mean tropopause heights (upper and lower) are indicated, and show a typical midlatitude behaviour. It is immediately apparent that the late summer months (January to March) are the least favourable for profiler performance, with SNRs less than -3 dB in the upper troposphere. During August and September, on the other hand, SNRs do not reach the -3 dB level until 16 km altitude. Additionally, it is worth noting that, because of the large enhancement at the tropopause, it is quite possible that the region above the tropopause may be observable, while the region below is not (*e.g.* November 1992 in Figure 4.10).

If we separate the data by the flight time, the only difference is an enhanced (by less than 3 dB) return in the morning (2300 UTC) flights during June-July and October-November in the upper troposphere. A full explanation for this observation is the subject of continuing investigation: in simplest terms, though, a decrease in

predicted returns in the upper troposphere will be due to a tendency towards nearer-to-adiabatic lapse rates (from Equation 2.16, as discussed in Section 2.4). It is the driving mechanism of such a tendency which requires solid identification. We note in passing that *Allen & Vincent* [1995], using the same set of AAP sonde data, report maxima in the seasonal gravity wave energy density through the same time and altitude regime; this has implications with regards to the r -exponent question, since a seasonal modification in gravity wave activity (and presumably turbulence) would impact on any assumptions of a given scattering model. This is consistent with the conclusions of *Nastrom et al.* [1985b], who demonstrated a seasonally-modulated diurnal variation in C_n^2 (which is closely related to M^2 , although on smaller scales than we are concerned with here) which they associated with upward-propagating gravity waves.

Extended comparisons with historical Buckland Park profiler data confirm the broad features of these estimates, although a direct comparison is complicated by changes in the profiler hardware and operating parameters over time. Nevertheless, we can readily infer a good comparison: the profiler data are consistent with colocated sonde launches (as seen throughout the SAFEx campaign), and the Buckland Park sonde data are consistent with the Adelaide Airport sonde data. For example, Figure 4.11 shows a number of comparisons between the two sonde stations and the profiler data. Note how significant features, such as the double tropopause in Flight 9, are reproduced by all three systems. Flight 10 has been included for comparison to this latter case: it is some three hours later than Flight 9, but has been plotted together with the AAP launch nearest the earlier SAFEx flight. The downward movement of the layer near 7.5 km is readily observable, as is the breakdown in the double-tropopause structure near 9 km. However, most attention should be paid to the generally excellent agreement between the AAP and SAFEx sonde launches, especially where the profiler differs (*e.g.* the lower stratosphere in Flights 9, 15 and 18); this serves to confirm our hypothesis that the large-scale measurements of the sonde systems are consistent, even when the modelling of the radar scattering mechanism (in essence, the r -exponent) may be less than perfect.

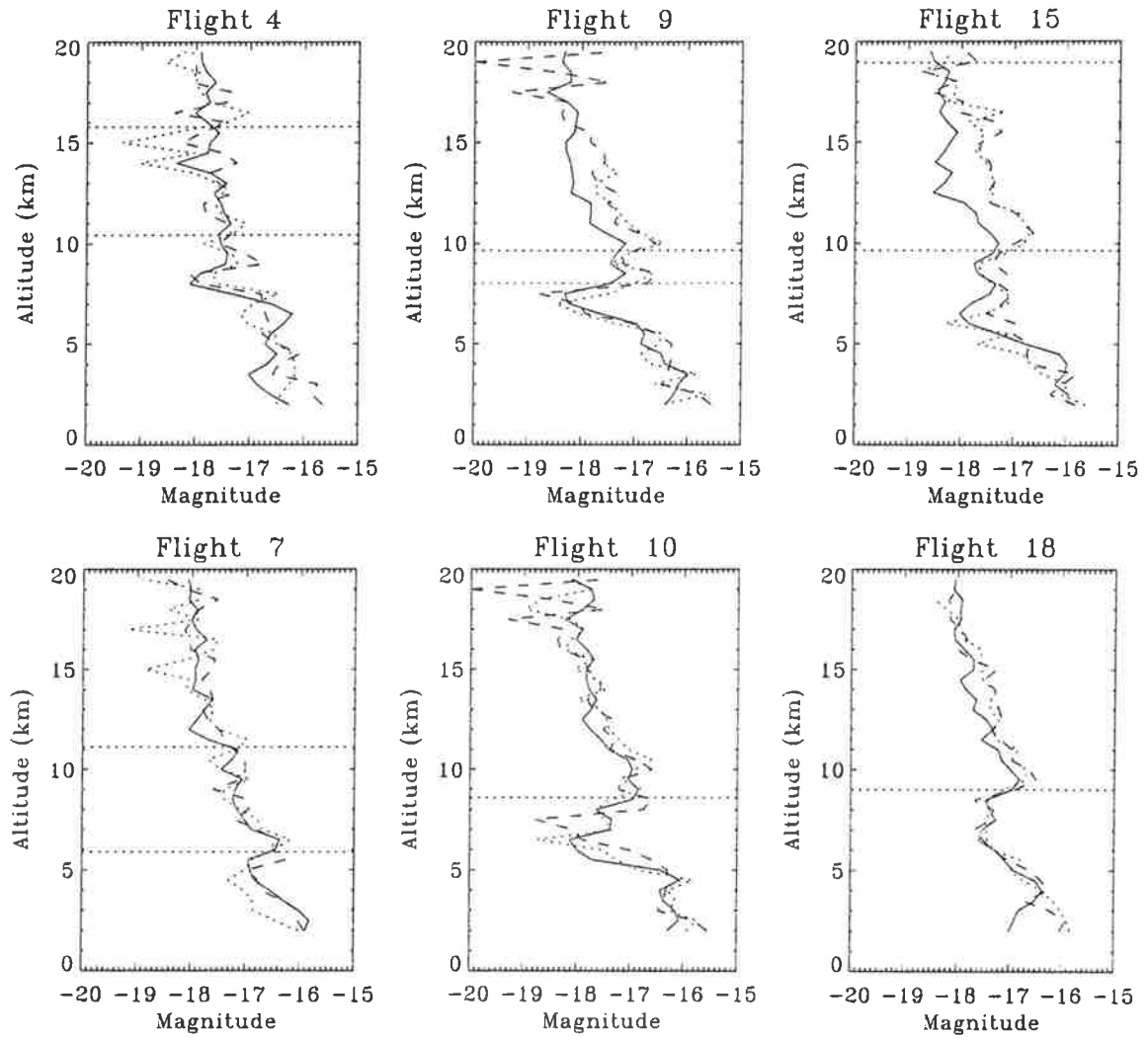


Figure 4.11: Comparison of AAP sondes with Buckland Park sondes and profiler data. The profiles are labelled by the SAFEx Phase IV sonde launch: solid, dotted and dashed lines are (respectively) range-corrected returned power ($r^2\mathcal{P}$), the full (wet) M^2 profile from the SAFEx flight, and the full M^2 profile calculated from the nearest AAP sonde launch (see Appendix C for exact timing details); units are log-scaled as in Figure 4.1. Dotted horizontal lines indicate SAFEx sonde-determined tropopause heights (the AAP determinations are near-exact matches, barring the near-fold Flight 7, where the tropopause identification is dubious in both cases). Only the radar profile has been shifted to match, as in Figure 4.1; the independent sonde launches have not been corrected in any way.

4.5.2 An Interlude: Predictive Considerations

Operationally, a SNR of 0 dB corresponds to the lowest level at which one could reasonably expect continuous height coverage; persistent biteouts are expected in regions with SNRs less than this. This is observed with the Buckland Park profiler, especially in February where useful data above 9 km are almost non-existent. The 0 dB contour can be used as a predictor of effective height coverage in this manner, with system changes modelled by shifting up and down contours. For example, we have recently doubled the transmitted power of the system (to 32 kW), corresponding to an expected increase of about 3 dB in SNR; in this case, the -3 dB contour would become the new “maximum height” indicator. Similarly, an identical system transmitting only 8 kW might use the 3 dB contour as a figure of merit.

As mentioned earlier, it is also important to bear in mind the details of the data collection. If we doubled the PRF, for example, we would essentially be doubling the transmitted power: that is, in a given amount of time, twice the original amount of energy would be put into the atmosphere. This would (in a perfect system) manifest itself as an approximately 3 dB increase in SNR over the data collected under the original set of parameters. Of course, there would be tradeoffs with regards to maximum height of returns and the observable velocity limit, as discussed in Section 2.2. In a similar manner, increasing the degree of coherent averaging will improve SNRs since the signal component coherently adds while the noise component incoherently cancels; however, this is only true while signal coherency can be assumed. Once signal coherency starts to break down, increasing the coherent averaging time fails to improve the SNR (and, in many cases, may actually decrease SNR as the “signal” component changes over time, and becomes confused with the background noise); frequency aliasing can be another unwanted result of excessive coherent averaging.

Physical changes to the array are another source of difference. This is perhaps most easily gauged by the array efficiency, although some attention needs to be paid to the actual beam (or polar diagram). Reducing system losses is an efficient way of increasing effective transmitted power; the Buckland Park profiler CoCo array efficiency is approximately 30%, and other array parameters are listed in Table 3.1. Note

4.5. PREDICTIONS

that changing the physical size of an aperture has little direct effect on SNRs: a larger array will produce a narrower beam, but this does not change the ratio of signal to noise within that beam provided that the scattering is consistent. Note, however, that a narrower beam serves to localise the transmitted and received energy and hence gives better directionality: in regions of anisotropic scatter, a narrower (vertical) beam will experience a relatively greater contribution from near-vertical layers or anisotropic scatterers, which may well improve SNR over a wider beam. Although this is a practical example, the experimenter generally has no influence over scattering mechanisms; however, it is a point which must be considered in the design and implementation of profiler systems.

Bearing these points in mind, we now move on to more predictive applications of profiler performance.

4.5.3 Darwin (12°S, 131°E)

Since 1989, a 49.8 MHz wind profiler has been operated by the Bureau of Meteorology at Darwin, in tropical northern Australia. Data from this profiler were available for June 1991 and January 1992, the latter being in the monsoon season. Hence, it is possible to compare our performance predictions with actual profiler data in these two periods, which are climatologically distinct. We shall start, however, by considering the variations in predictions over a full twelve month span. Sonde launches from Darwin were made at 0830 LST and 2030 LST (Darwin and Adelaide are at the same UTC, and both experience the daylight savings launch-time effect outlined previously).

Figure 4.12 shows a selection of monthly-average $\overline{M^2}$ profiles together with the associated variances. A number of features are worthy of note: $\overline{M^2}$ decreases at about 2 dB km^{-1} in the stratosphere, with a geophysical variation of about 1 dB (standard error in the mean). The tropopausal enhancement near 17 km is of the order of 18 dB; note that a single high-altitude tropopause is characteristic of tropical regions. Careful examination of the tropospheric portion of the profile reveals significant differences between the monsoon (November to March) and non-monsoon periods. The distinct enhancement near 2 km seen in the climatologically dry winter months corresponds to

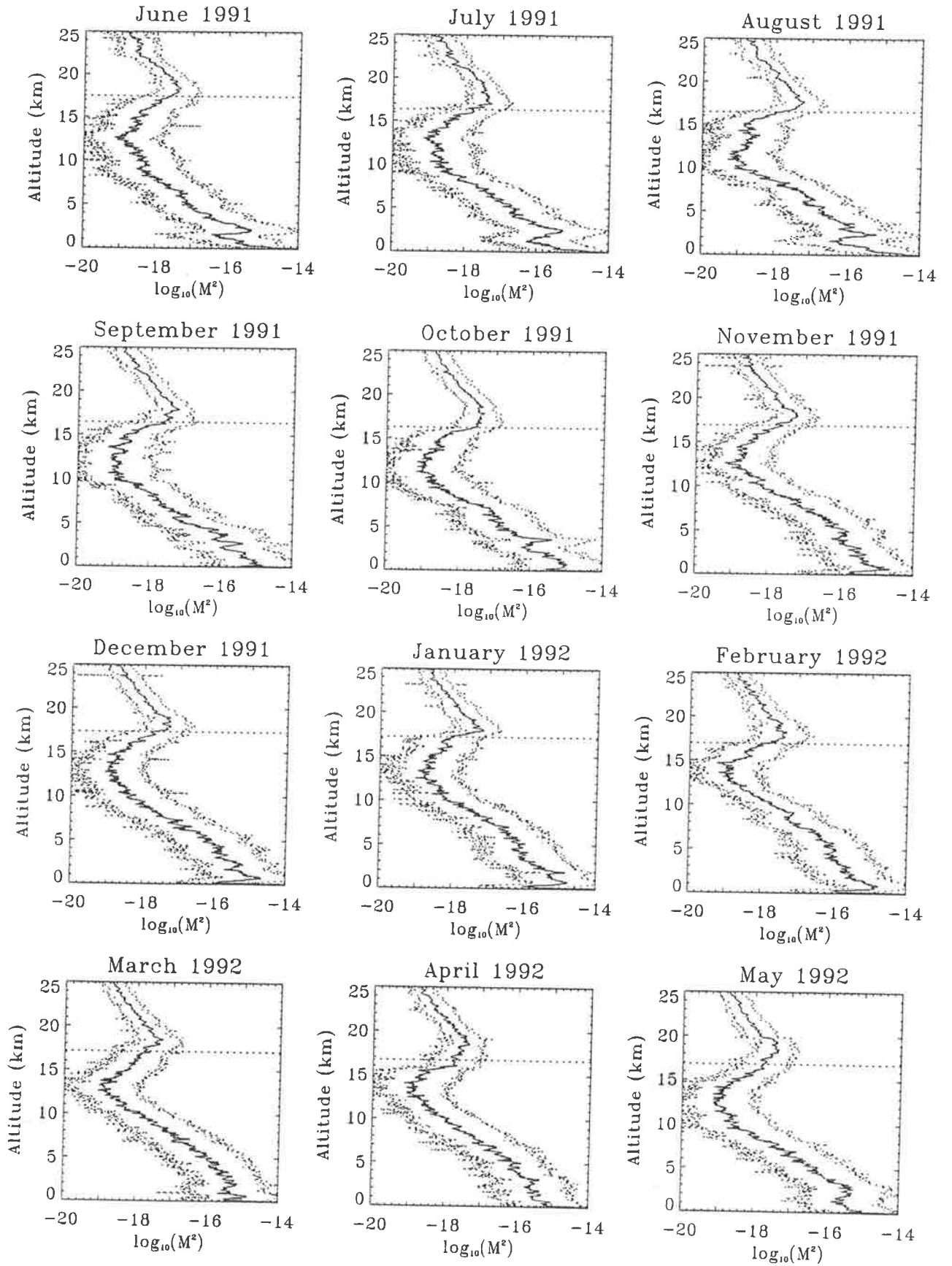


Figure 4.12: As in Figure 4.9, but for Darwin sonde data; approximately sixty sonde flights contribute to each plot (two flights per day, at 0830 LST and 2030 LST).

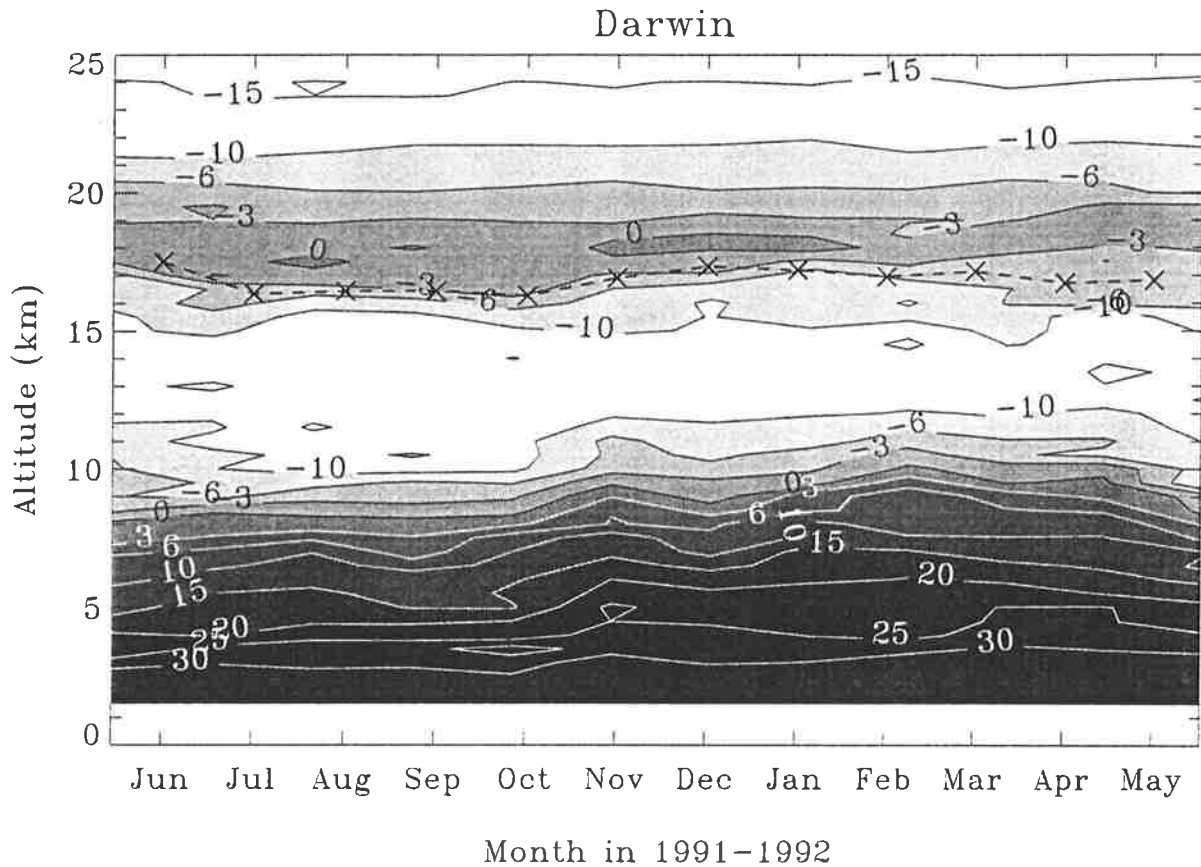


Figure 4.13: Predictions of observed SNR, derived from Darwin sonde data. As in Figure 4.10, but dashed lines and $\times$'s indicate single-cluster average tropopause levels.

a marked decrease in atmospheric water content above that level (recall from Equation 2.16 that M^2 is proportional to q and $\partial \ln q / \partial z$). The summer monsoon period is the wet season, with higher water vapour throughout the troposphere, peaking near 5 km due to convective redistribution. Hence, the summer months show a change in slope near 5 km: about 2.5 dB km^{-1} below and 5 dB km^{-1} above. The $\overline{M^2}$ slope during the winter months is closer to 4 dB km^{-1} throughout the low-mid troposphere, while the variation is about 2 dB regardless of season. Note that this behaviour is different to midlatitude sites, where the maximum in convective activity tends to occur around the winter months.

Figure 4.13 combines the predicted SNRs from the monthly-mean $\overline{M^2}$ profiles into a shaded contour plot, assuming a profiler identical to the Buckland Park VHF system. Sonde data from morning and evening flights have been combined in this figure. The most noticeable feature is the improved performance through the summer monsoon

period, mainly due to enhanced levels of water vapour content in the mid-troposphere. The lower stratosphere also displays an enhancement during this time, possibly due to increased gravity wave activity from convective sources in the troposphere. *Allen & Vincent* [1995], for example, analyse data from the nearby Gove (12°S, 136°E) station, and find a maximum in temperature variance (which can be related to gravity wave energy density; see *Kitamura & Hirota* [1989], for example) in the summer lower-stratosphere. They also observe a smaller maximum at the same altitude in August, where we see another small enhancement in predicted SNR.

If the data are separated by time of flight, we see a marked difference between the morning (Figure 4.14) and evening (Figure 4.15) launches. The morning flights generally predict stronger SNRs in the upper troposphere and lower stratosphere. As a general note, local effects are important in the tropics, and we have not considered wind-driven effects at all: for example, in regions where air is moist and unstable, air of maritime origin flowing on-shore (a “sea breeze”) may lead to increased vertical motion, resulting in afternoon showers or thunderstorms (convection) which can affect the upper troposphere. For example, the adiabatic cooling of rising air may enhance a tendency for temperature profiles in the upper troposphere to approach the adiabatic lapse rate, at which point M^2 tends towards zero (again, from Equation 2.16) and returned powers would be expected to decrease. An in-depth study of temperature tendencies as related to the occurrence of afternoon convection is beyond the scope of this thesis, although it does present an interesting avenue for further research. As we are interested in average performance, however, these differences will for the moment be considered as a contribution to the variance in predicted performance.

The discussion so far has involved only the Darwin sonde data. We can use the available Darwin profiler data to confirm some of the predictions, giving us an independent verification of our approach. To begin with, it must be noted that while the Darwin profiler is similar to the Buckland Park system (they are both near-50 MHz CoCo-array profilers), there are a number of differences in the specific setups and parameters. For example, the noise level is approximately 15 dB between 8 km and 25 km (less at lower altitudes due to the removal of clutter from the raw Darwin data), some 5 dB higher than the Buckland Park profiler. Note that only the SNR is an absolute

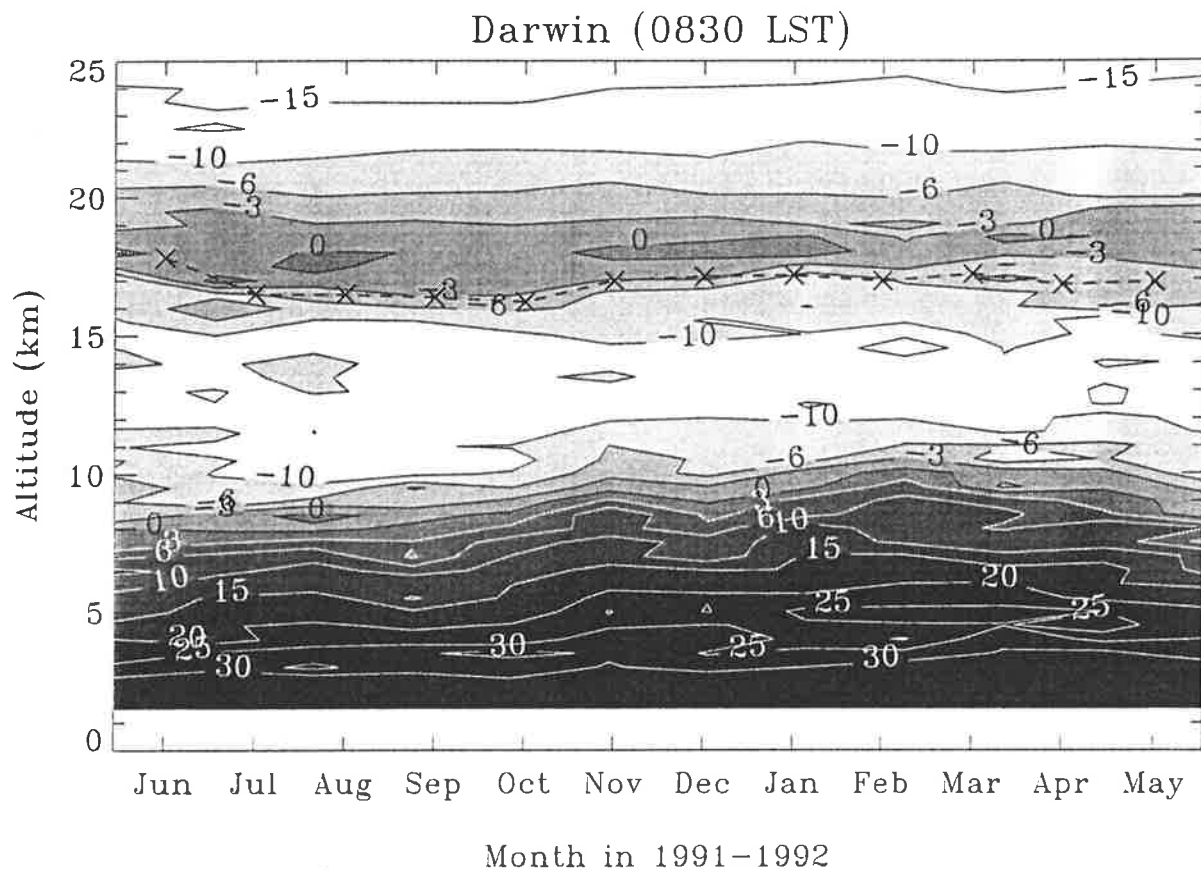


Figure 4.14: As in Figure 4.13, but for Darwin 0830 LST sonde data.

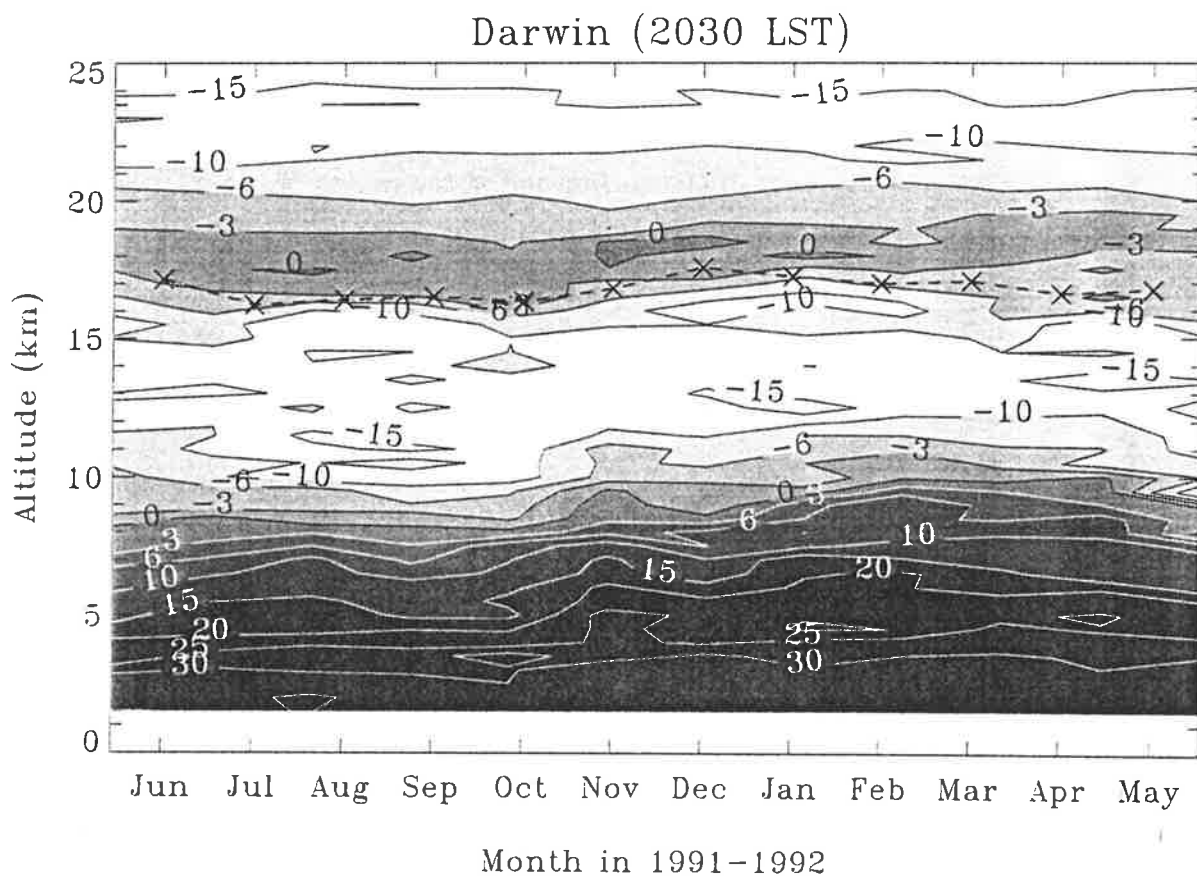


Figure 4.15: As in Figure 4.13, but for Darwin 2030 LST sonde data.

measurement for uncalibrated profilers: the returned power and noise level estimates are generally arbitrarily defined depending on the details of time-series analysis. If we follow through Equations 4.4 and 4.5 with this correction, we see that our estimates of SNR will be approximately 5 dB too low, relative to the Buckland Park system. The Darwin profiler has a peak power output of 45 kW (versus 16 kW at Buckland Park), giving another 5 dB or so in (absolute) SNR. Hence, we would expect to have to add some 10 dB to the predicted SNR values to make a direct comparison between the Adelaide and Darwin systems. We would also expect some differences due to specific system-related effects (receiver efficiency, for example), which we do not have the ability to measure; however, they are unlikely to exceed another 3 dB or so.

Figure 4.16 shows the predicted SNR for June 1991, based on the twice-daily Darwin sonde launches. The predictions have been corrected by 12 dB to enable a more direct comparison with the profiler data, as outlined above. A number of features are worth noting in this more detailed graphic, as an aid towards comparison: generally low SNRs in the upper troposphere over the first six days; the sharp step between 8 km and 10 km on June 8; the enhancement (and underlying “tongue” or reduced SNR) at 11 km between June 14 and June 19; the “double tongue” enhanced SNR feature between 10 km and 15 km between June 19 and June 28; the altitude and magnitude of the tropopausal enhancement above 17 km; and the series of wavelike striations in the tropopausal enhancement in the second half of the month. In general terms, the tropopause is predicted at a level of about 9 dB, over an upper-troposphere minimum of about -3 dB.

The corresponding Darwin profiler observations are displayed in Figure 4.17. They have been matched to the time resolution of the predictions by averaging for one hour either side of each sonde flight. This produces a somewhat smoother graphic than the preceding figure, and moderates the extrema (raising the minima and lowering the maxima), but is a far more representative comparison than would be achieved using the unsmoothed profiler data. The height resolutions have already been matched in the prediction process, via the resampled running-mean technique described earlier. Finally, the same contour levels and shading have been used in the two Figures.

The comparison is remarkably good, with every feature of the predictions described

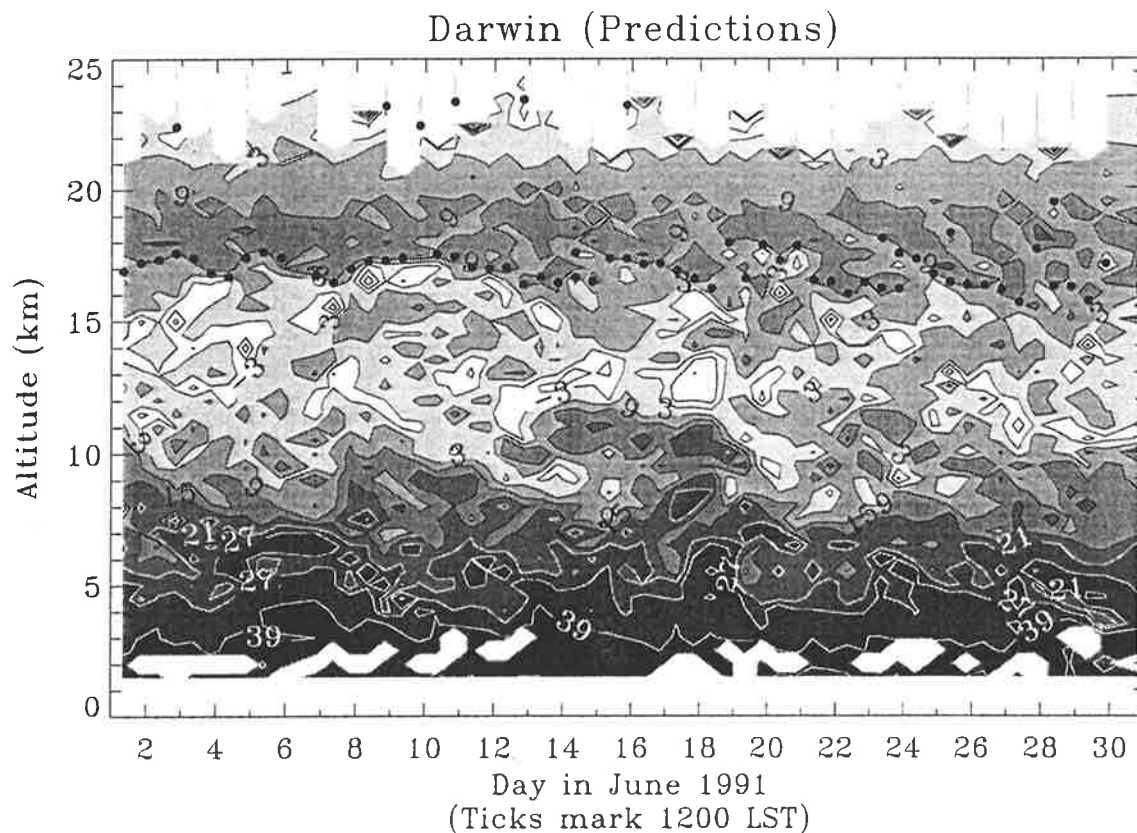


Figure 4.16: Predicted SNR for June 1991, from Darwin sonde data. A correction of 12 dB has been applied. Contour levels are at 6 dB intervals from -15 dB to 39 dB. Solid dots mark individual sonde tropopause retrievals.

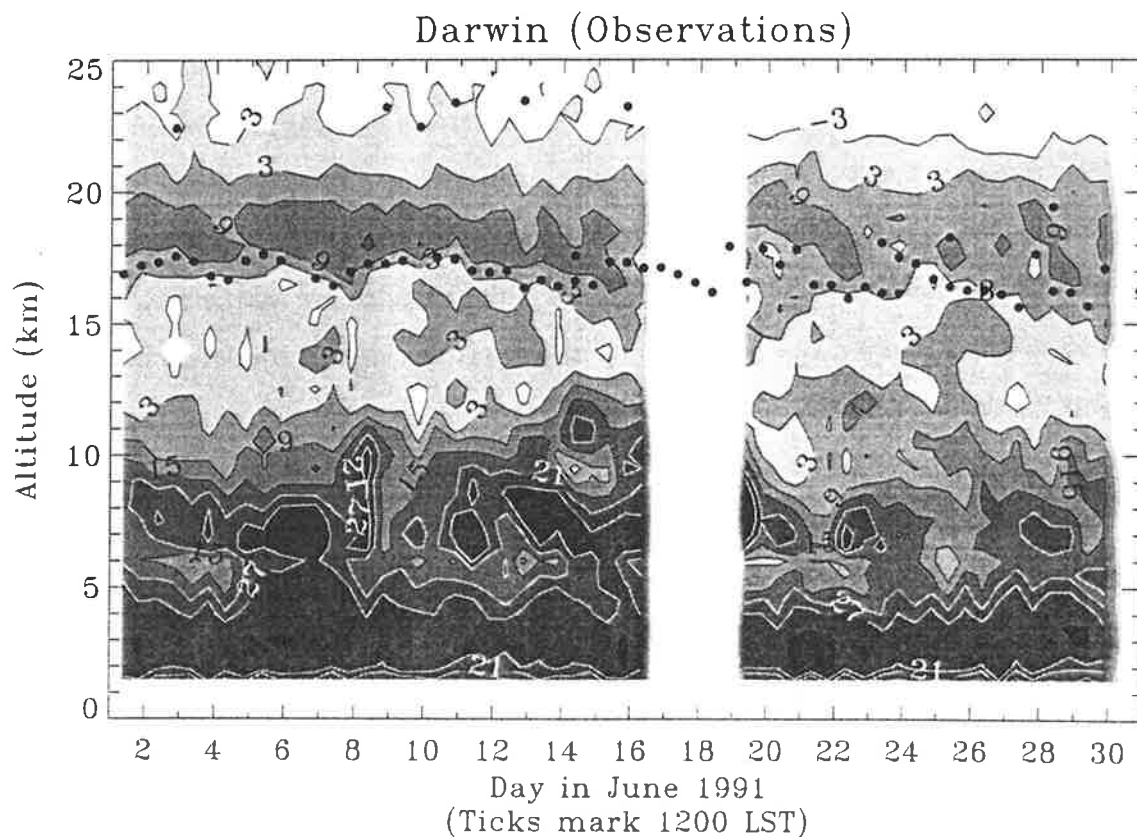


Figure 4.17: Observed SNR for June 1991, from Darwin profiler data. Contour levels, shading, and tropopause marks are identical to Figure 4.16. The profiler was inactive for about two days around June 18.

above reproduced in the observations. The tropopausal striations are not as pronounced in the observations, although there is evidence of such behaviour between June 20 and June 22 (and perhaps around June 28). However, this is a result of the sensitivity of the contouring process, and a marginal difference in the magnitudes of the enhancement between the observations and predictions: if a contour level of 6 dB is used for the observations, the striations are clearly visible. The generally raised minimum level of the observations (throughout the upper troposphere) is mainly due to the effects of averaging. As a final point, note how the altitude of predicted and observed SNR levels in the lower stratosphere are in excellent agreement: the 3 dB level lying between 20 km and 21 km, for example.

Figures 4.18 and 4.19 show, respectively, the predicted and observed SNRs at Darwin for January 1992. Key features are again reproduced: the general increase in SNR in the upper troposphere across the periods January 3–6, 9–13 and 21–27, for example. The comparison is not as good as that seen in June, especially in the low-mid troposphere, due to the monsoon conditions in January. The enhanced layer observed above 5 km in particular is not seen in the predictions. However, the upper troposphere and lower stratosphere are modelled quite well.

The difference between the non-monsoon (June) and monsoon (January) cases can perhaps be seen more clearly in Figure 4.20, where the average predicted SNRs have been compared with the mean observations (averaged, as above, for one hour either side of each sonde flight). The June profile matches quite well, although the enhancement above 5 km is not well modelled; the January observations are markedly higher in the low-mid troposphere. Another point of interest is in the altitude levels of the major features (the upper troposphere minimum and the tropopausal enhancement): the observations appear to be shifted by some 500 m upwards relative to the predictions. After checking, a mis-calibration in the Darwin profiler range gates has been confirmed (P. May, personal communication, 1995); this indicates an additional application of the technique, as a means of calibrating profiler returns as a function of range (typically an difficult problem, due to time delays in the transmission and reception systems being interpreted as range delays). Although one would shy away from making a single-profile comparison of the altitude of features due to the possible motion of layers over

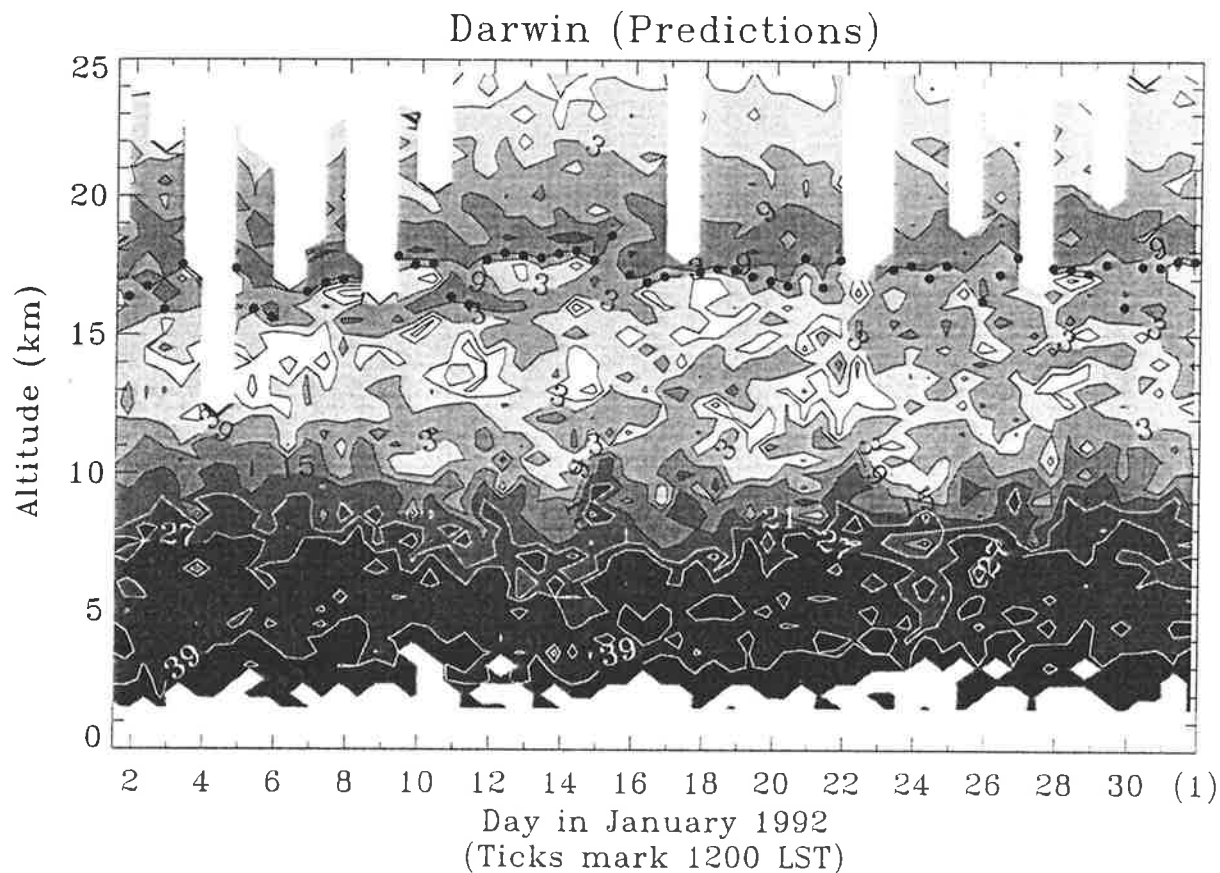


Figure 4.18: As in Figure 4.16, but for January 1992.

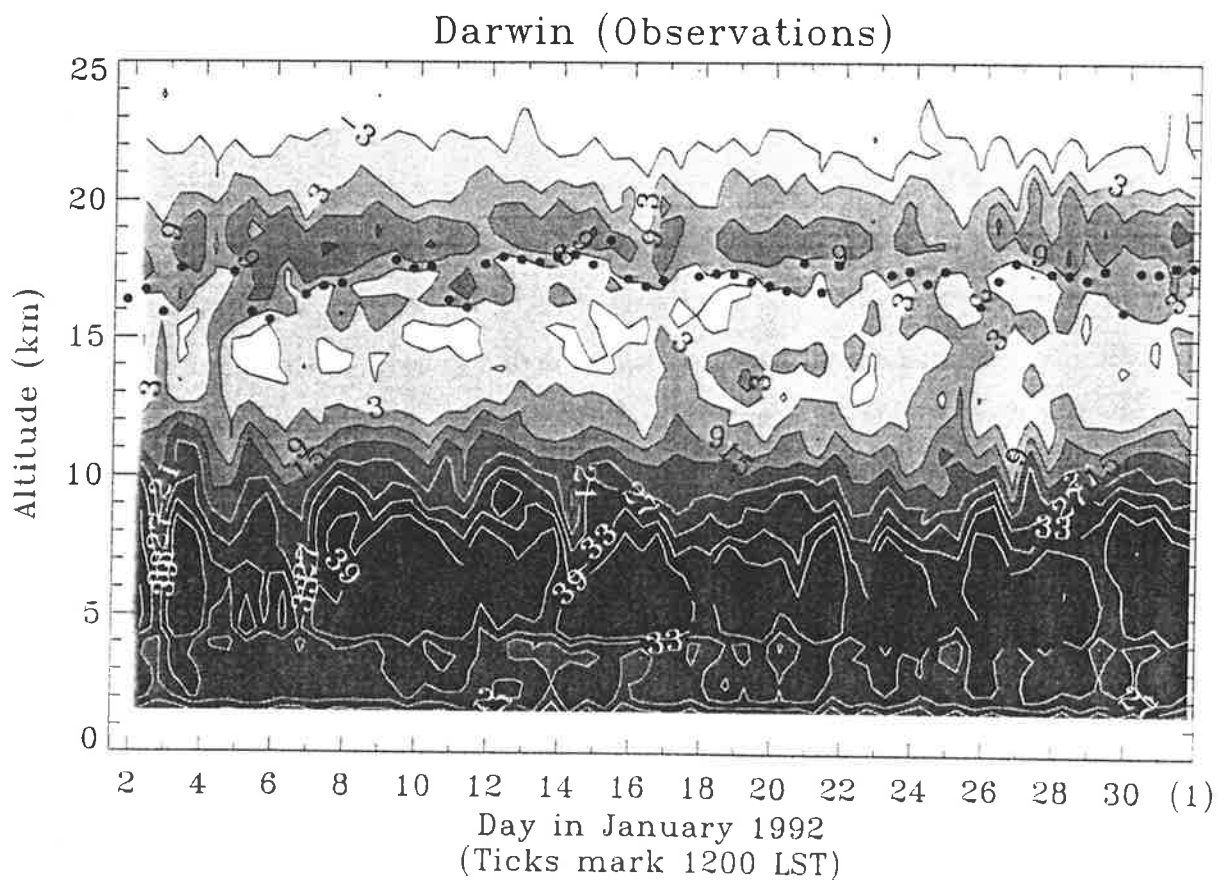


Figure 4.19: As in Figure 4.17, but for January 1992.

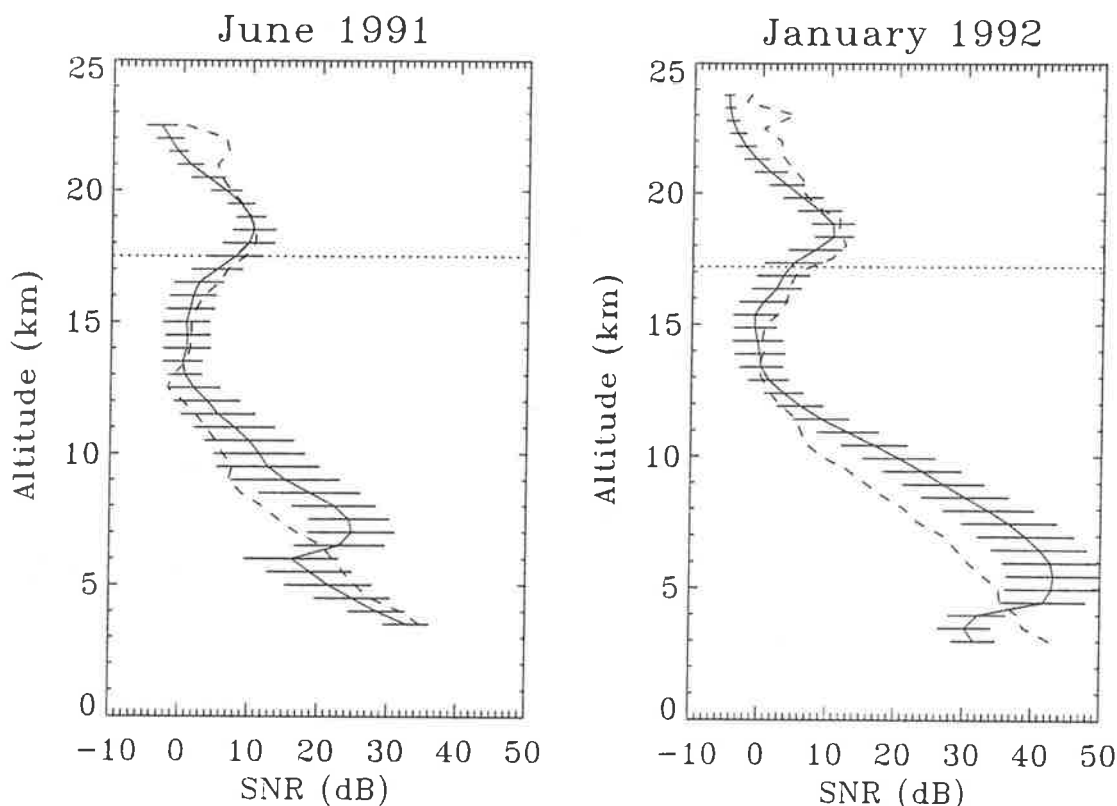


Figure 4.20: A comparison of average predicted (dashed) and observed (solid) SNR profiles at Darwin for June 1991 and January 1992 (12 dB correction applied to the predictions; see text). Error bars on the observations show one standard deviation either side of the mean. The average tropopause height is indicated by a horizontal dotted line.

the period of a few hours, and layer slopes due to mesoscale effects, such differences should average out over the month or so that we are examining here.

In an effort to explain the significant differences between the June and January profiles, we examine a number of individual flights in some detail. Figure 4.21 shows data from three flights in June 1991. Enhancements due to temperature inversions (frame (a), near 9 km) and moisture (frame (b) near 6 km, and frame (c) near 10 km) are obvious, and appear in both predictions and observations. However, the major enhancement near 7 km in frame (c) is not well modelled; similar enhancements are observed throughout the month's data. We believe this to be an water-phase echo effect: note the location of the freezing (or melting) level, at 0°C. Most moisture above this level will be in the form of ice which, as it falls through the melting layer, can become covered with a layer of liquid moisture. The radar brightband effect near

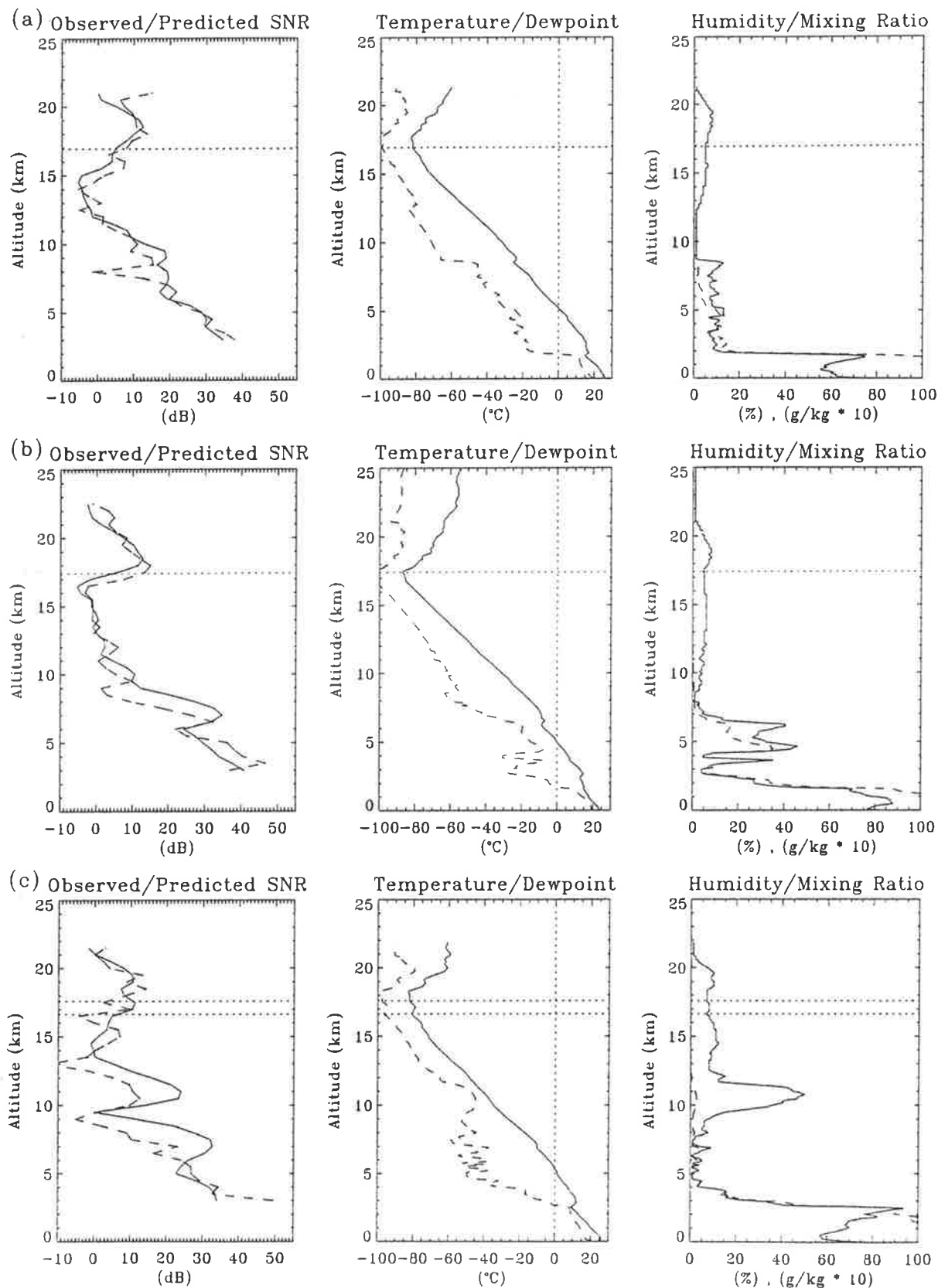


Figure 4.21: Each frame of three diagrams shows the observed (solid) and predicted (dashed) SNR (dB); the temperature (solid) and dewpoint (dashed) in Kelvin; and the relative humidity (solid, %) and mixing ratio (dashed, $\text{g kg}^{-1} \times 10$). Horizontal dotted lines indicate sonde-determined tropopause retrievals. (a) 2030 LST 1 June 1991, (b) 0830 LST 5 June 1991, (c) 2030 LST 14 June 1991.

the freezing level is well-documented (see, for example, *Rajopadhyaya et al.* [1994]) when moisture and freezing temperatures occur together. This is not modelled by our estimation procedure, but with our interests mainly reserved to the upper troposphere, we do not regard it as a significant problem for determining profiler performance.

Note that in June, Darwin is climatologically dry above the boundary layer; in the monsoon season, however, the lower atmosphere is dominated by moist effects. Figure 4.22 shows three typical flights from January 1992, and the elevated moisture content is obvious. This significantly enhances the freezing-level returns, and leads to the general underestimation seen in the SNR predictions. The freezing level return is dominant near 6 km or so, and the higher moisture content leads to a higher contribution to the observed radar returns due to freezing and melting related effects. It is worth noting that the freezing process releases latent heat which, together with the vertical motions associated with tropical convective processes, can contribute to the temperature inversions observed above the freezing level [*May & Rajopadhyaya*, 1994]; such temperature inversions are evident in both of the previous Figures.

As a final point in the discussion of the Darwin data, we note again the tropopausal striations in predicted (and, to a lesser degree, observed) SNR in the second half of June 1991. These wave-like features have a period of about five days, and extend over about 5 km in the lower stratosphere. These are characteristic of tropical waves (rather than internal gravity waves), and further investigation into their strong signal in the $\overline{M^2}$ profiles and feedback into the observed tropopause heights is certainly warranted. Unfortunately, wind data from the profiler is not available for this time, which limits investigation of this case; however, there is evidence of similar signals in other months at Darwin. One possibility is tropical-extratropical interaction: that is, we could be seeing the tropical response to transient features such as fronts (G. Holland and P. May, personal communication 1994). This is the subject of ongoing research.

4.5.4 Alice Springs (23°S, 134°E)

Twelve continuous months of data from Alice Springs, in the central Australian desert region, were analysed. Only one radiosonde launch was made each day, near 2300 UTC

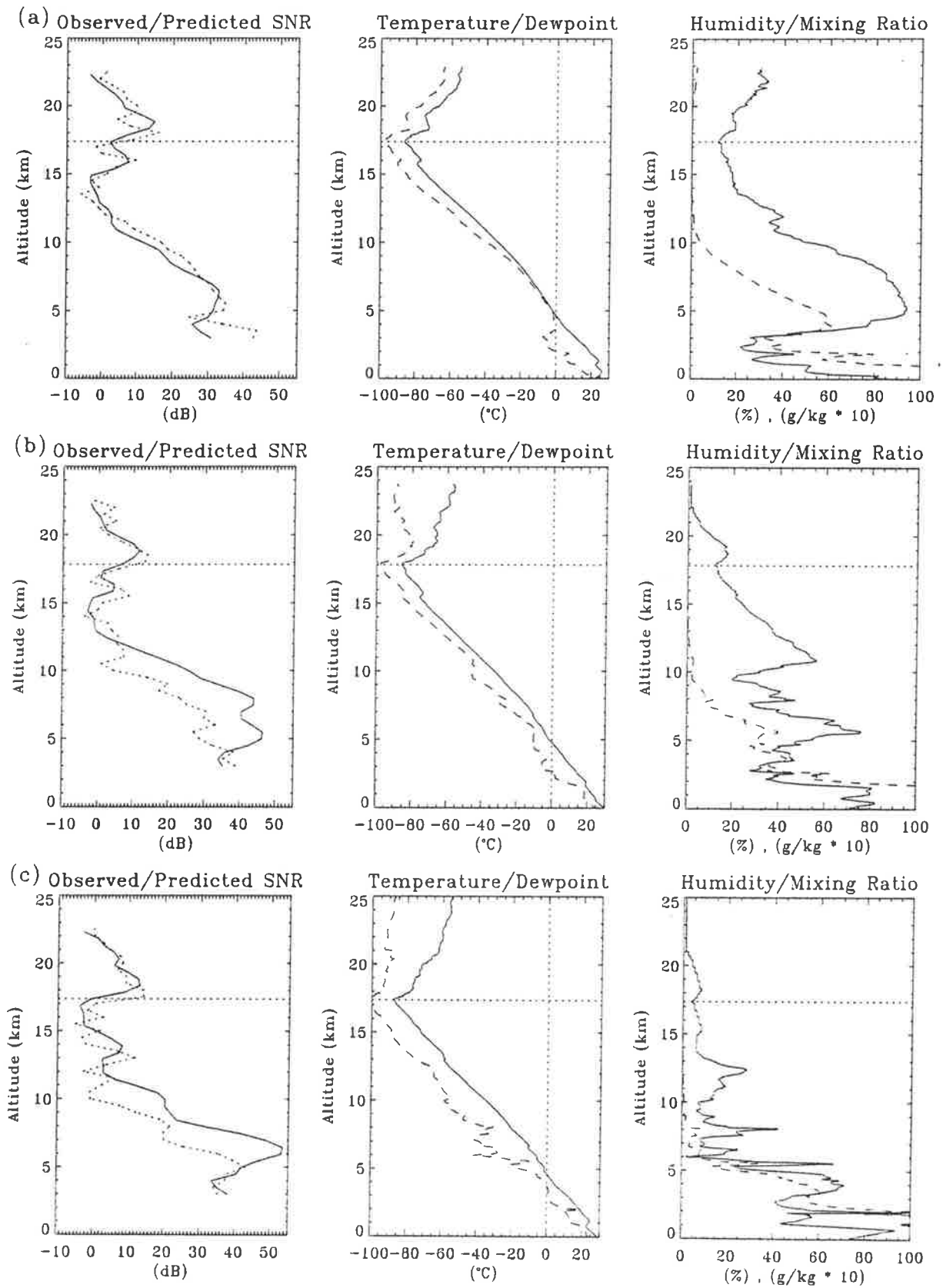


Figure 4.22: As in Figure 4.21, but for: (a) 0830 LST 4 January 1992, (b) 2030 LST 9 January 1992, (c) 0830 LST 18 January 1992.

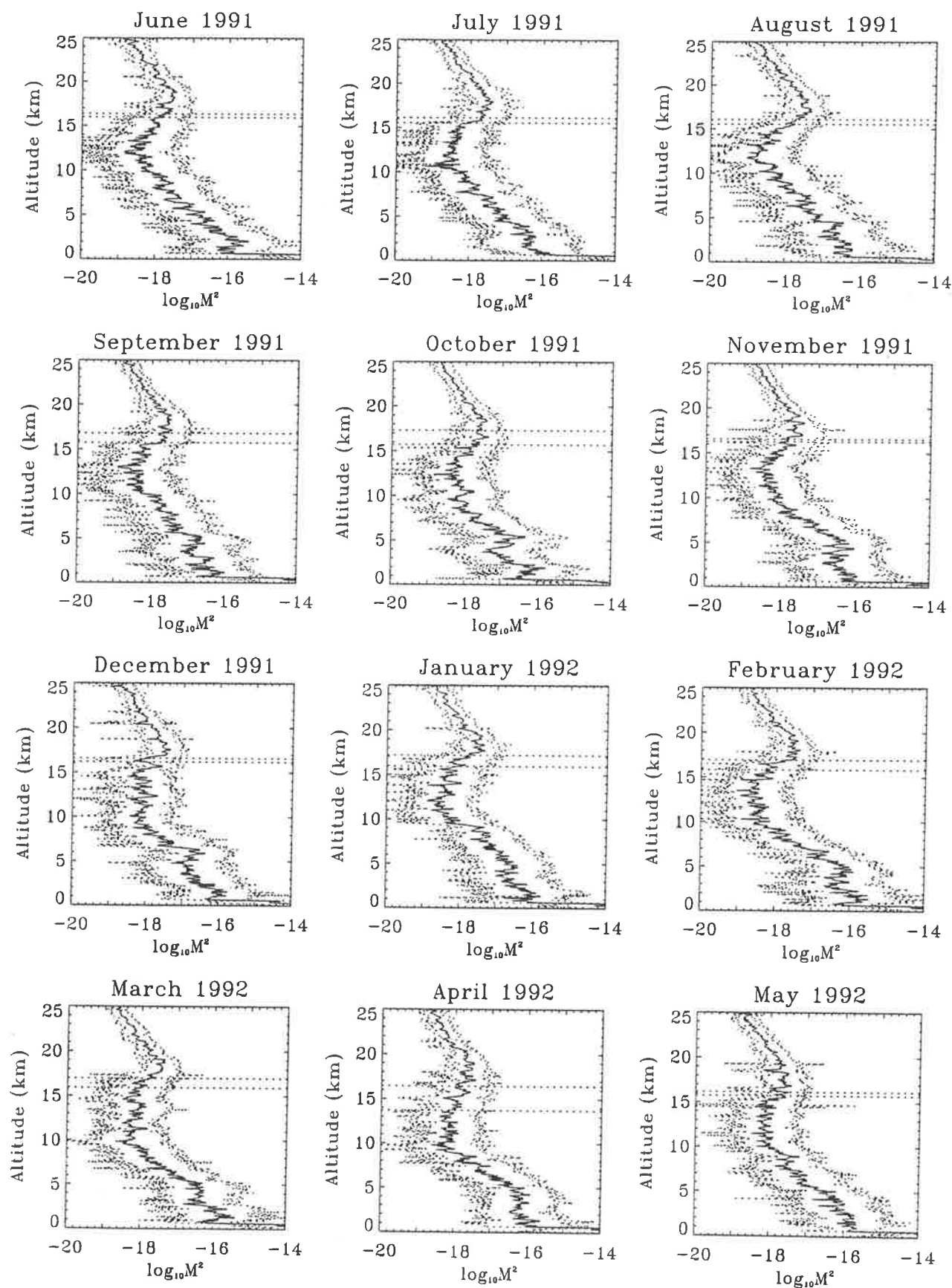


Figure 4.23: As in Figure 4.9, but for Alice Springs sonde data; approximately thirty sonde flights contribute to each plot (one flight per day, at 2300 UTC/0830 LST).

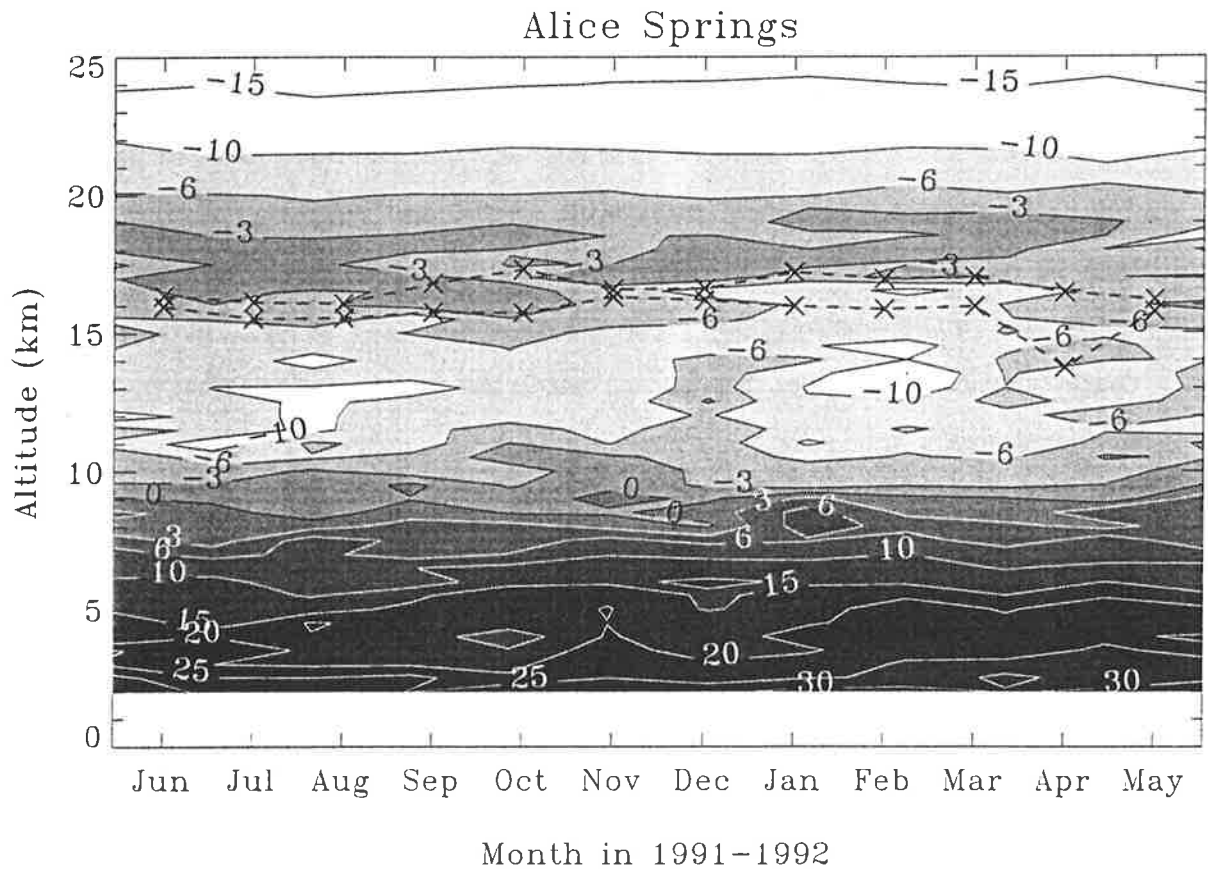


Figure 4.24: As in Figure 4.10, but derived from Alice Springs sonde data.

(0830 LST). This presents some limitations on our predictions since no night data available, but any differences between night and day should only affect the boundary layer (although we note the small differences observed in the Adelaide predictions). The monthly mean $\overline{M^2}$ profiles (a selection of which are displayed in Figure 4.23) show some common features: most noticeably, a near 20 dB enhancement in the boundary layer, due to the remnants of a regular surface radiation inversion in the lowest few hundred metres of the atmosphere; and a smaller enhancement around 5 km, associated with moisture gradients. $\overline{M^2}$ decreases at about 1.3 dB km^{-1} through the stratosphere, and the geophysical variation in absolute values again ranges from about 2 dB in the troposphere down to 1 dB in the stratosphere.

The tropopause at Alice Springs shows very little seasonal variation, lying between 15.5 km and 16.5 km throughout the year. The predicted SNR contours shown in Figure 4.24, however, display a significant variation in the upper troposphere—nearly 10 dB difference between the minima in August and December at around 13 km. The

drop-off in SNR is rapid in the lower troposphere (around 5 dB km^{-1} up to 8 km), but slower in the upper troposphere. It is apparent from the figure that if an operational profiler is to reach the tropopause on a regular basis, attention must be paid to overcoming the upper-troposphere minima in August and February. If this is done, in fact, an Adelaide-type profiler should obtain returns on a regular basis up to 23 km or above, due to the relatively slow stratospheric decrease. This again indicates that the characteristics of the upper troposphere can be the major limiting factor in operational profiler design.

4.5.5 Mawson (67°S, 63°E)

Radiosonde launches from Mawson Base in Antarctica were only available for May through to December 1993, on a twelve-hourly basis with flights at approximately 0000 UTC and 1200 UTC each day. If the data are grouped by sonde launch time, differences in the analysed fields are small and almost exclusively restricted to the boundary layer. Note that at this latitude, differences in incoming solar radiation are generally more dependent on season than time of day. The average monthly $\overline{M^2}$ profiles show a number of features which can be seen in the example Figure 4.25. The local maximum near 2 km altitude is due to a consistently-appearing layer of enhanced moisture and a seasonally-dependent temperature inversion, and is about 10 dB stronger in winter than summer: although the moisture gradients are similar between the seasons, the temperature inversion near 2 km altitude is much stronger in winter. The low polar tropopause near 10 km marks another local maximum, although the enhancement is significantly less than the corresponding features seen at lower latitudes; it also extends over a smaller vertical range. It is worth noting once again the results reported by *Allen & Vincent* [1995] for the nearby Davis station (69°S, 78°E), which show a minimum in gravity wave energy density around August to October near 8 km, the same region where we see a minimum in the predicted profiler SNR.

Above the tropopause, $\overline{M^2}$ decreases at about 1.3 dB km^{-1} in winter, although the summer profiles tend to a convex rather than linear shape; that is, a higher rate of decrease (nearly 2 dB km^{-1}) in the summer upper stratosphere. The geophysical

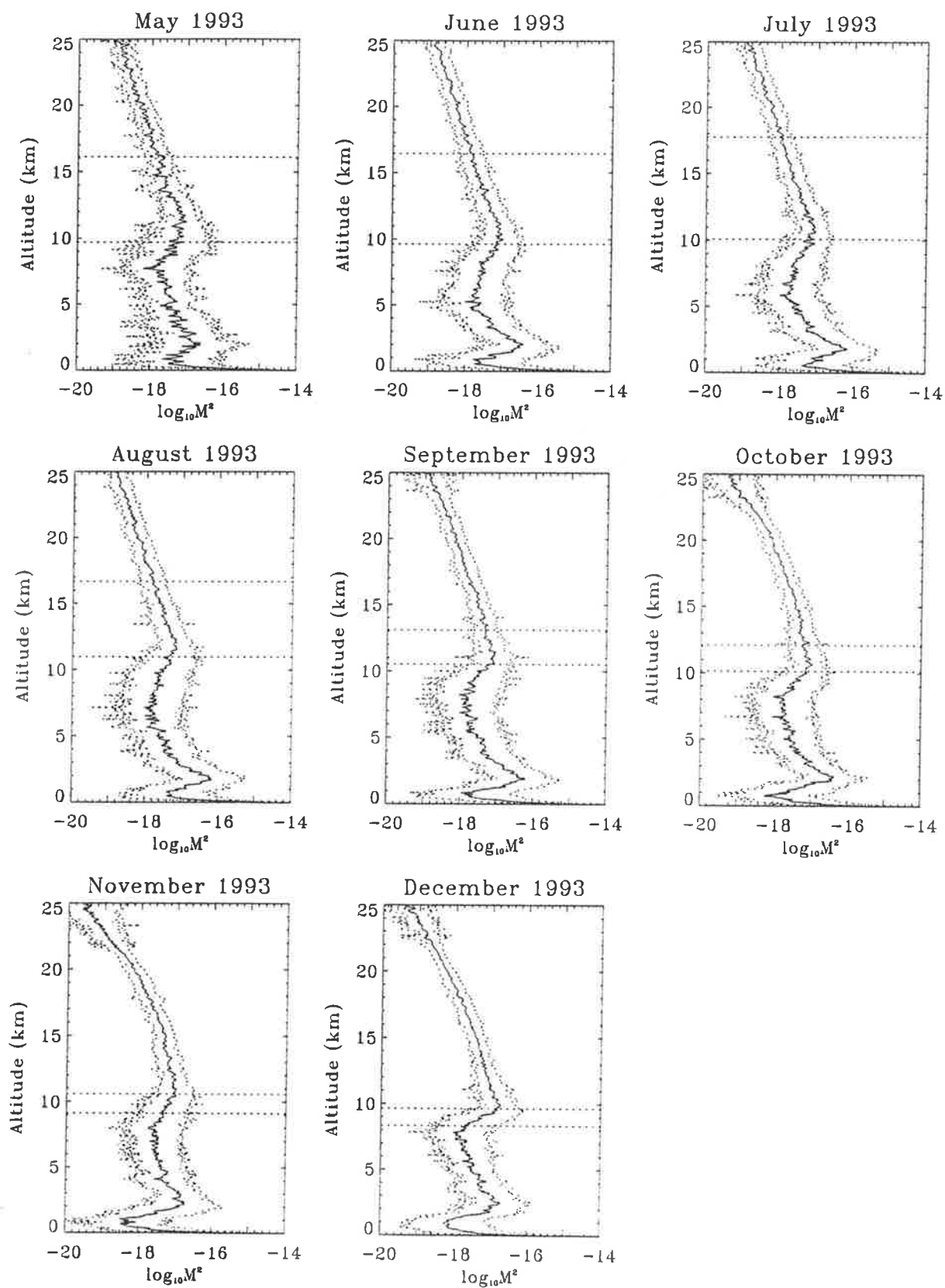


Figure 4.25: As in Figure 4.9, but for Mawson sonde data: approximately sixty sonde flights contribute to each plot (two flights per day, at 0000 UTC and 1200 UTC).

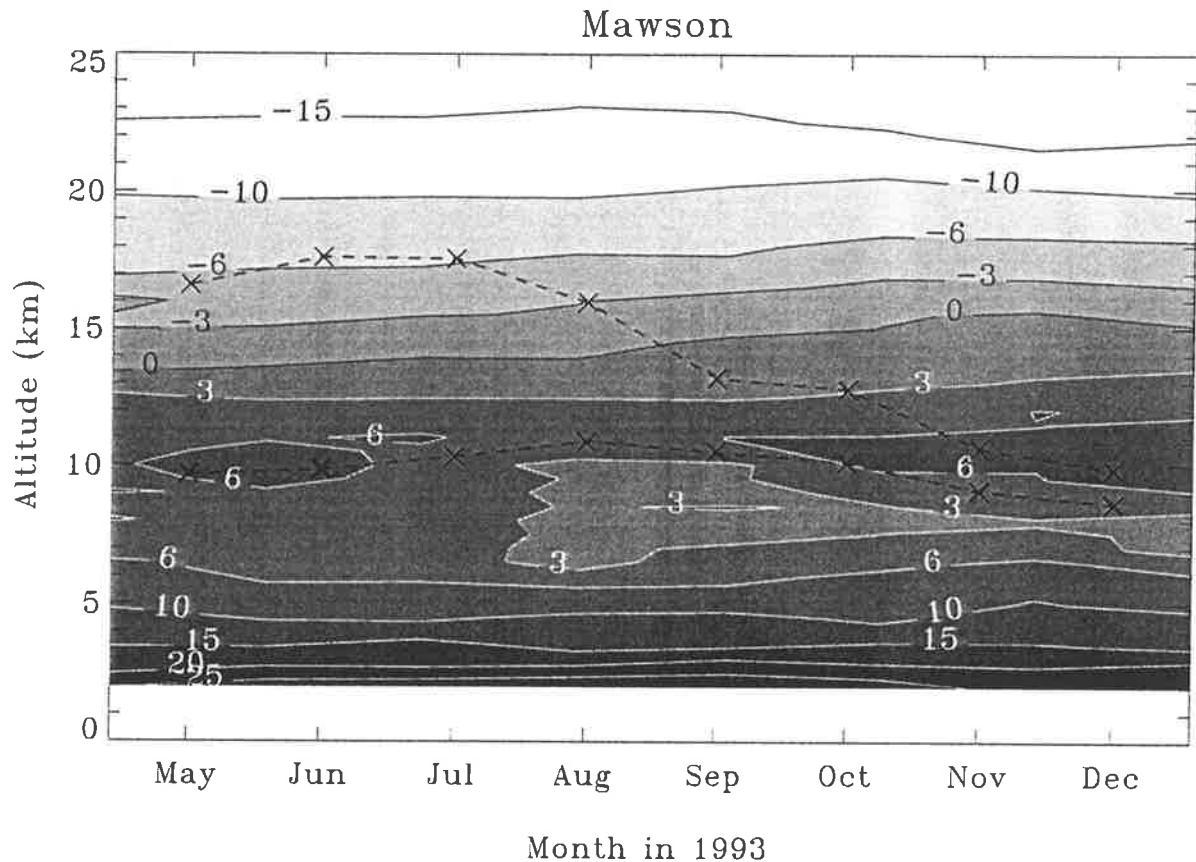


Figure 4.26: As in Figure 4.10, but derived from Mawson sonde data.

variation is again approximately 2 dB in the troposphere, and 1 dB in the stratosphere.

The predicted SNR graphic Figure 4.26 shows very good performance over the entire analysed region, with the tropospheric minimum being greater than 0 dB throughout. The minimum occurs around the September equinox, but the gradients are generally smooth in height and time. Notice how the lower tropopause follows the base of the SNR enhancement. There is no noticeable enhancement associated with the sonde-determined upper tropopause, but this feature is quite weak in the thermodynamic profiles. We note the report of *Balsley et al.* [1993] on the installation of a 50 MHz wind profiler for the study of tropospheric and mesospheric winds in Antarctica; the results are awaited with interest.

4.6 Summary and Conclusions

The extensive comparisons at Adelaide and Darwin show that our method of estimating the altitude coverage of VHF vertical-beam profiling systems from climatological radiosonde (p, T, q) data is generally applicable. This is the first time that qualitative—let alone quantitative—predictions of profiler performance from independent sources have been made, and has significant implications for the installation of future networks of wind profiling radars. Historical radiosonde data provides baseline information from which key profiler design criteria (such as average power) can be properly estimated, rather than guessed.

However, there are of necessity a number of limitations to such a simple model. In general, the lowest 5 km of the atmosphere is dominated by moisture effects, and $\overline{M^2}$ displays a decrease with altitude through the troposphere which is nearly constant across a variety of latitudes and climates. Similarly, the behaviour in the lower stratosphere is independent of location, and is dominated by fluctuations in potential temperature gradients. The tropopause marks the boundary between these two regimes, and the magnitude of $\overline{M^2}$ there is near-constant since it is strongly dependent on dT/dz which is always near-zero there. As the tropopause altitude varies with latitude, the height of the enhancement varies, making the characteristics of the upper troposphere the limiting factor in profiler performance.

Local effects can have a significant effect: directly, through moisture variations or temperature inversions producing localised enhanced returns; or indirectly by gravity wave generation and propagation leading to enhanced potential temperature fluctuations in the upper troposphere and lower stratosphere. In addition, the model does not take account of ice effects, so that a strong radar brightband near the freezing level is not seen in the predictions. However, as far as the model is intended to perform (monthly averages, for example), it behaves well. An additional line of enquiry from comparisons of predicted and observed power returns, it should be possible to estimate average height profiles of scatterer nature (via the exponent of the range dependence). Comparisons with the seasonally-variable gravity wave energy density at different locations provide avenues for further joint radar-radiosonde investigations of the nature

of the lower middle-atmosphere.

One point which we have not mentioned in our previous discussion involves varying background noise levels. By averaging background levels over days or months, we make predictions based on an average, but this does not take account of effects like the diurnal variation in skynoise seen in Adelaide (and Darwin) due to the passage of the galactic plane over the radar. This can raise noise levels by up to 25 dB for a couple of hours each day, resulting in an oscillatory pattern in returned signal power and SNR. Again, this is a local effect which can only be evaluated on a case-by-case basis.

One of the key assumptions of the model is a scattering profile with height which goes as r^{-2} at all altitudes. Local variations in gravity wave activity—such as we would expect in Adelaide during winter due to the increased level of frontal activity, in Darwin during the monsoon, or near the jet stream exit around the tropopause—could affect the r -exponent in the short term, as would any seasonal variation in the nature of the scatterers. These variations would be expected to occur over height and time; looking at the difference between model-predicted profiles and observations would be one way of examining such variation in the atmospheric radiowave scatterers. However, attention would need to be paid to competing effects, such as increased turbulence leading to enhanced volume-scatter, but also breaking down any stably-stratified layers which produce enhanced vertical-beam returns under Fresnel scatter. That is, increased turbulence could lead to either an increase or a decrease in returned power, depending on the degree of returns due to reflection-like processes from stable layers.

We have examined the performance of profilers employing a vertical beam (vertical Doppler beams, or spaced antenna systems, for example); the question now arises as to the performance characteristics of UHF or off-vertical VHF profilers. Noting that returned power in these cases is proportional to C_n^2 from Equation 2.6. *VanZandt et al.* [1978] give the following relationship for thin layers:

$$C_n^2 = a^2 \alpha' L_o^{4/3} M^2 \quad (4.6)$$

where a^2 is a universal constant close to 2.8 and α' is a ratio of eddy diffusivities

usually taken equal to unity. However, this requires that the value of M^2 is due to turbulent mixing, and not the bulk thermodynamic quantities as we have evaluated from the radiosonde profiles in the work described above.

This viewpoint is supported by our observations, where the bulk $\overline{M^2}$ profiles show a marked enhancement at the tropopause. Tropopausal enhancements are a feature of returns from vertical VHF beams (*e.g.* Gage & Green [1979]), but not in well off-vertical VHF or UHF returns, where turbulent backscatter rather than Fresnel scatter dominates. Enhancements near the tropopause in the latter cases are generally associated with upper-level jets, where increased wind speed produces increased turbulence, increased C_n^2 , and hence an elevated level of backscatter [Nastrom *et al.*, 1986]; or with increased gravity wave activity due to the change in N^2 at the tropopause [VanZandt & Fritts, 1989].

Although our method may predict off-vertical VHF performance with some degree of accuracy in the mid- to upper-troposphere, where the scattering mechanisms are similar, we would not expect it to do so at higher levels. A natural extension of this work is to examine the performance of UHF and off-vertical VHF beams; this would require a similar approach, but involve statistics on the small-scale fluctuations in M^2 due to turbulence rather than the bulk approach we have taken here.

Chapter 5

RADAR OBSERVATIONS OF THE TROPOPAUSE

To be sure of hitting the target, shoot first, and call whatever you hit the target.

— *Ashleigh Brilliant*

The tropopause marks the boundary between the troposphere and the overlying stable stratosphere. The change in temperature lapse rate and the associated variation in atmospheric stability between these two regions manifests itself in enhanced reflections of vertically incident VHF radio waves. Knowledge of the tropopause altitude is of importance in fields such as stratosphere-troposphere exchange (STE), upper-level frontogenesis, and temperature soundings derived from satellite measurements of radiance. The aim of this chapter is to discuss a method for detecting the height of the tropopause on a real-time basis using a VHF radar, and extend it to a study of multiple layers. We present a number of interesting case studies, although detailed examinations of the background meteorological situation are restricted to Chapters 6 and 7.

5.1 The Fresnel Scatter Model

In the mid-1960's, *Atlas et al.* [1966a] reported monostatic radar detection of the tropopause using powerful non-Doppler radars. The observed echoes were attributed to turbulent scattering from inhomogeneities in clear-air refractive index; these turbulent echoes, although found more often at or near the tropopause than elsewhere, were weak and sporadic. This was to be expected—the radars used were sensitive to length scales of the order of 10 cm (wavelengths near 1 GHz) and, if we exclude birds and insects, turbulence is the primary mechanism responsible for radar echoes at these scales. However, at the larger wavelengths to which VHF radars are sensitive, partial specular reflection becomes increasingly important when the radar is directed vertically at stable horizontal laminae of refractive index.

In fact, only a few of the observed features of the enhanced specular echoes are consistent with either the mechanisms of anisotropic turbulent scattering or partial reflection. The observed features can be summarised as follows:

- the most intense echoes are associated with very stable atmospheric layers, typically located in the lower stratosphere;
- the echoes are very anisotropic and appear specular, since the magnitude decreases appreciably as the zenith angle of the probing beam increases;
- autocovariance functions are typically very wide—the echoes fade much slower than observed for turbulent echoes;
- the wavelength and pulse-length dependences are generally not characteristic of a few-layer partial reflection hypothesis.

In order to produce a more satisfactory model of the specular echoes, the Fresnel scattering approach was introduced (as described in Section 2.4). A key component of this model is the assumption that the scattering medium is coherent in the two dimensions transverse to the probing wave, and random in the dimension parallel to the wave direction. In other words, we assume horizontal stratification of refractive index such that $n(z)$ varies randomly in the vertical—think of the atmosphere as a

series of thin slabs, each with slightly different n . Transverse coherence can extend from a few wavelengths to well over a Fresnel zone. At VHF wavelengths, this corresponds to contributing irregularities of a few hundred metres in horizontal extent and of the order of $\lambda/2$ thick.

Strong transverse coherence occurs, for example, in hydrostatically stable regions of the atmosphere, where vertical motions are suppressed and small-scale motions take on a two-dimensional character. The tropopause region is particularly suited to this type of model, as the large stability gradients there serve to reduce cross-tropopause fluxes. When viewed with a vertically directed radar, one then sees enhanced echoes at the location of (stable) horizontally layered structures due to in-phase scattering from the half-wavelength Fourier component of the stratified irregularity structure parallel to the wave direction. An off-vertical beam suffers partial cancellation upon scattering, and has a much reduced magnitude.

5.2 Tropopause Altitude Determination

In this section, an algorithm is developed for the purpose of tropopause height detection. It is based on the Fresnel scattering model described above and in Section 2.4, and is an extension of the ideas of *Gage & Green* [1982a] and derivative work [*Riddle et al.*, 1983; *Green & Gage*, 1984; *Riddle et al.*, 1984; *Gage et al.*, 1986; *Sweezy & Westwater*, 1986].

Combining Equations 2.7 and 2.21 gives the radar equation for Fresnel scatter:

$$\mathcal{P}(r) = \frac{\alpha^2 \mathcal{P}_t A_e^2}{4\lambda^2 r^2} F(2k, r)^2 \Delta r \overline{M^2}/4 \quad (5.1)$$

If we now recall the “dry” equation 2.18 for the generalised gradient of potential refractive index, M_d , and make the assumption that the upper troposphere is dry enough to neglect the humidity contribution to M , then we can write:

$$\mathcal{P}(r) = \beta \left[\frac{p}{T^2} \left(\frac{dT}{dz} + \Gamma_d \right) \right]^2 f(r) \quad (5.2)$$

where β is a constant incorporating all of the radar and reflection constants, and $f(r)$ is the range dependence of the backscattered echoes from the radar equation, which

is usually assumed to go as r^{-2} . As we have discussed in Chapter 4, variations in the exponent of r are not uncommon, especially in regions of enhanced stability or specularly.

Following *Gage & Green* [1982a], define a range-normalised received power $\mathcal{R}(r)$ by:

$$\mathcal{R}(r) = \frac{\mathcal{P}(r)}{f(r)} \equiv \mathcal{P}(r)r^2 \quad (5.3)$$

to remove the effects of range dependence. We can also integrate the hydrostatic equation (1.3) to give:

$$p = p_0 \exp \left(- \int_{r_0}^r \frac{dz}{H} \right) \quad (5.4)$$

where $p_0 \equiv p(r_0)$ is the pressure at some reference level r_0 , and H is the (temperature-dependent) scale height of the atmosphere:

$$H = p/\rho g = RT/\bar{m}g \approx (29.3)T \quad (5.5)$$

($R = 8.3143 \text{ J K}^{-1} \text{ mol}^{-1}$, $\bar{m} \sim 0.028964 \text{ kg mol}^{-1}$; T in Kelvin, H in metres).

Then, substituting for pressure p and taking the natural logarithm:

$$\ln(\mathcal{R}(r)) = \ln \left(\frac{\beta p_0^2}{T^4} \left[\frac{dT}{dz} + \Gamma_d \right]^2 \right) - 2 \left(\frac{r - r_0}{H} \right) \quad (5.6)$$

and hence:

$$\ln(\mathcal{R}(r)) + 2 \left(\frac{r - r_0}{H} \right) = \ln \left(\frac{\beta p_0^2}{T^4} \right) + 2 \ln \left(\frac{dT}{dz} + \Gamma_d \right) \quad (5.7)$$

Denote the left-hand side of this equation by $\mathcal{Q}(r)$, which is obviously another normalised returned power, corrected for the exponential decrease of pressure with altitude [*Riddle et al.*, 1983]:

$$\mathcal{Q}(r) = \ln(\mathcal{R}(r)) + 2 \left(\frac{r - r_0}{H} \right) \quad (5.8)$$

The Bureau of Meteorology uses the standard World Meteorological Organisation (WMO) definition to locate the tropopause from aerological diagrams:

The tropopause is located at the lowest altitude (above the 500 mb level) above which the temperature lapse rate does not exceed 2 K km^{-1} for at least two kilometres.

Near the tropopause, we expect to find a level where $dT/dz \sim 0$; in general, this will be above the tropopause. This leads us to define $\mathcal{Q}_0$ as the value of $\mathcal{Q}(r)$ at the altitude where the temperature gradient vanishes. That is:

$$\mathcal{Q}_0 \equiv \mathcal{Q}|_{\frac{dT}{dz}=0} = \ln \left(\frac{\beta p_0^2}{T^4} \right) + 2 \ln (\Gamma_d) \quad (5.9)$$

and hence we can rearrange Equation 5.8 in terms of $\mathcal{Q}_0$:

$$\mathcal{Q}(r) = \mathcal{Q}_0 + 2 \ln \left(\frac{1}{\Gamma_d} \frac{dT}{dz} + 1 \right) \quad (5.10)$$

If we identify the “radar tropopause height” with a temperature lapse rate of 2 K km^{-1} (*i.e.* $dT/dz = -2 \text{ K km}^{-1}$) and recall that the dry adiabatic lapse rate Γ_d has a numerical value of about 9.8 K km^{-1} (from Equation 1.7) then:

$$\mathcal{Q}(r_{\text{trop}}) \approx \mathcal{Q}_0 - 0.46 \quad (5.11)$$

where r_{trop} is the range from the radar to the tropopause; that is, the tropopause altitude. Note that this is a correction to Equation 5 in *Sweezy & Westwater* [1986], which suffered from some confusion between natural and base-10 logarithms. If the procedure above (based on Equation 5.8) is carried out with base-10 logarithms, then the resultant constant in Equation 5.11 is 0.20 rather than 0.46.

If we know the atmospheric thermodynamic parameters at the time of the radar sounding, as well as the radar parameters, we can substitute into Equation 5.9 to find the required value of $\mathcal{Q}_0$. This is not a practical solution for two reasons: firstly, we wish to determine the height of the tropopause from the radar data alone (after all, if we had radiosonde soundings of the atmosphere giving the pressure and temperature profiles, we could get the tropopause height directly); and secondly, many operational profilers (including the Buckland Park radar) are not fully calibrated, so the radar parameters are unknown. The alternative is to recall that we defined $\mathcal{Q}_0$ as the value of $\mathcal{Q}$ where dT/dz was zero. We would expect this to hold for some region above the tropopause, into the stratosphere (especially in Adelaide, as we discuss in the next section); hence, we can approximate $\mathcal{Q}_0$ by some average value of $\mathcal{Q}$ above the tropopause.

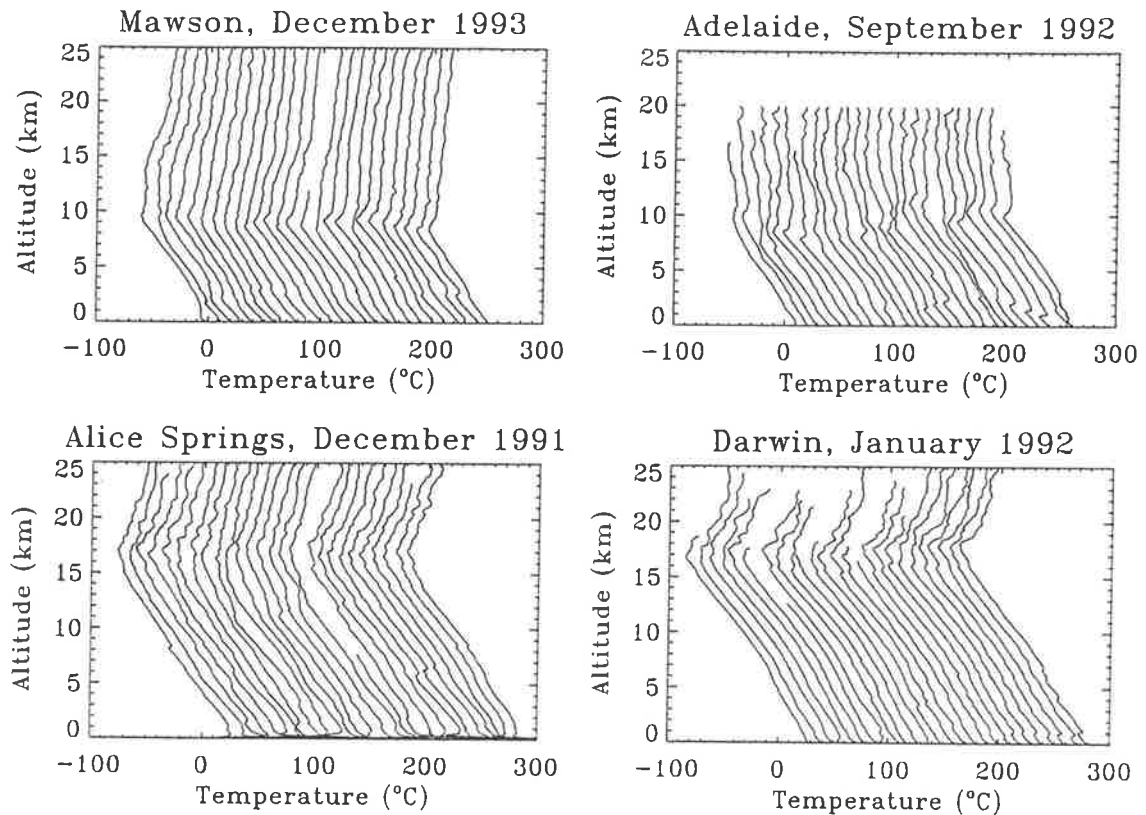


Figure 5.1: Example sequential temperature profiles from four sites. Successive profiles are displaced by 10°C, and are about twelve hours apart in time. The change in temperature gradient marks the local tropopause.

5.3 Comparison with Climatology

The initial justification of this method requires confirmation of the relationship between returned power and $\overline{M^2}$. The arguments and evidence presented in Chapter 4 can be applied here in precisely the same manner, and will not be repeated. However, it is worth commenting on two points. Firstly, the question of the form of the range dependence. In most cases, we found that r^2 was satisfactory, although at times exponents approaching 2.4 were better, especially in cases where specular scatterers dominated over turbulent scatter.

Secondly, the nature of the atmosphere over Adelaide is typical of midlatitude sites, with both a “polar” and “tropical” tropopause evident through most of the year (the polar tropopause dominates in winter, while the tropical signal dominates in summer). Typical altitude profiles of temperature show a region of near-isothermal

temperature between these two layers near 10 km and 15 km. This is in contrast to a tropical site such as Darwin, where a single tropopause near 16–17 km marks the boundary between regions of nearly monotonic temperature decrease and increase with height; or, conversely, to a polar site such as Mawson where the single tropopause is near 10 km, at least during the southern hemisphere summer. Some examples of this behaviour are displayed in Figure 5.1. These differences are important in the practical implementation of a tropopause altitude determining algorithm which makes use of characteristics of the atmosphere *above* the tropopause. We shall examine data from Darwin and compare it with Adelaide later in this chapter.

A number of sample $Q(r)$ profiles are displayed in Figure 5.2. The profile labelled (1) shows an example of a high tropopause, while profiles (6)–(13) show a sequence of eight evaluations over thirty hours. These profiles are taken from SAFEx IOP IV/3, which is detailed in Chapter 6 (see also Appendix C for the actual sonde flight times). The average $Q(r)$ profile plotted is calculated directly from the hourly average returned power; the individual $Q(r)$ values calculated from each of the twelve returned power profiles are indicated by dots at each altitude level. Note that the resolution of Q is rather coarse, since the returned powers are only recorded to the nearest dB (and $Q(r)$ here is calculated post-collection). However, the relatively small spread in Q values is obvious, with around half the contributing values overlapping. The variance over longer periods, however, is obvious. In anticipation of the implementation of the tropopause detection algorithm, we display the radar-derived tropopause heights and the corresponding sonde retrievals.

The $Q(r)$ profiles have a number of common features. Firstly, the similarity with returned power is readily apparent (compare, for example, Figure 4.1). Since $Q(r)$ is essentially a pressure- and range-normalised returned power, this is not surprising! The effects of the normalisation are most obvious in the stratosphere, where $Q(r)$ is more constant with altitude than either $\mathcal{P}(r)$ or $\overline{M^2}$. Ideally, the algorithm requires a constant value of $Q(r)$ for some distance above the tropopause. We see that this is not always the case: similarly to the discussion in Chapter 4, variations in the range-dependence exponent are the most likely cause (compare the relevant panels in Figure 4.1, and the discussion in Section 4.4). Since any errors will involve

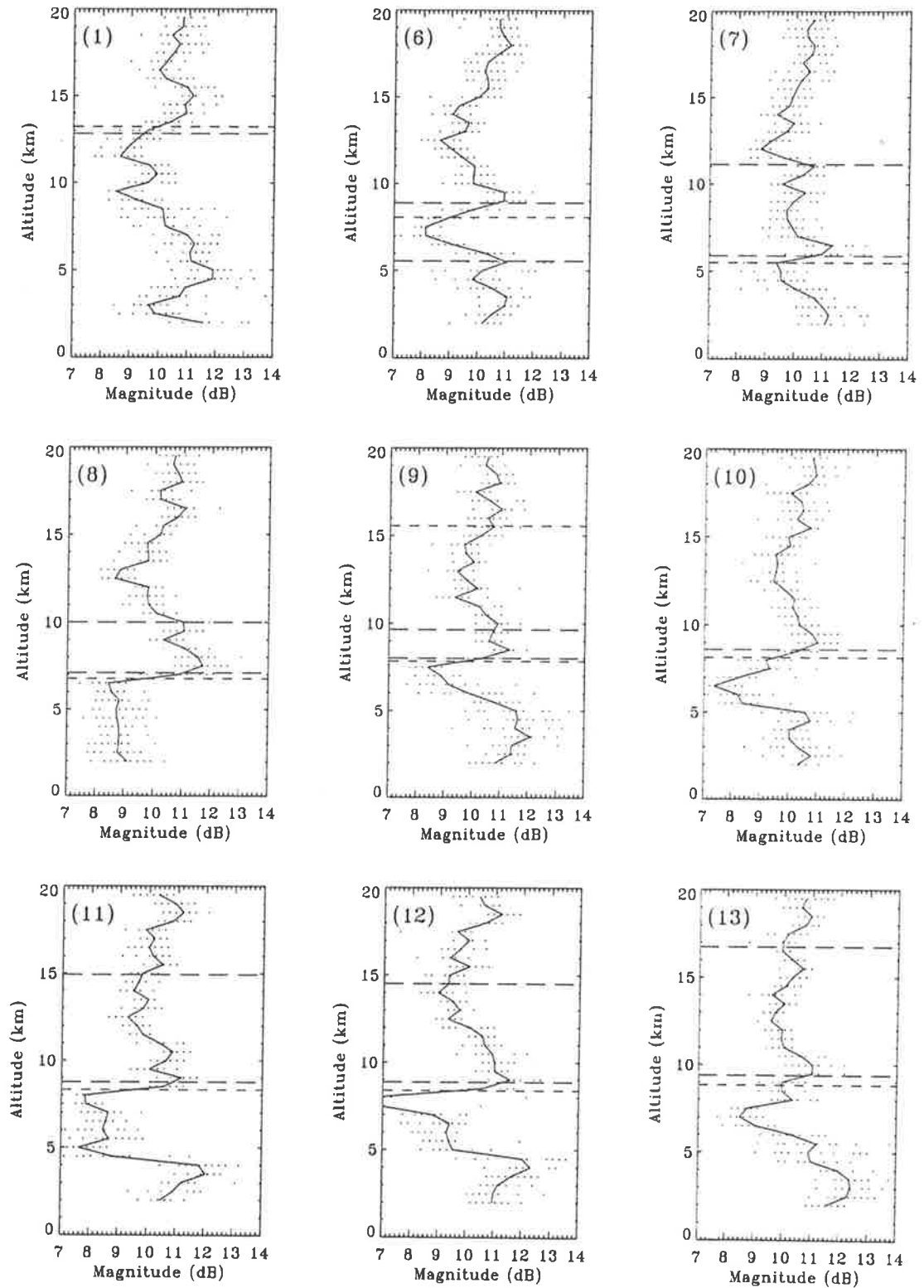


Figure 5.2: Example $Q(r)$ profiles from SAFEx IOP IV/3. The profiles (solid) are evaluated from the hourly-average returned power around the time of the sonde flight listed in the top left corner of each plot. The dots show the spread in the twelve individual five-minute profiles contributing to the hourly average. Tropopause heights determined from the profiler data are short-dashed lines, while the sonde-determined heights are long-dashed lines.

underestimation (rather than overestimation) of the range-dependence exponent, the effect on the algorithm will generally be an underestimation of Q_0 , and a resultant underestimation of $Q(r_{\text{trop}})$ and r_{trop} (since $Q(r)$ increases from a minimum below the tropopause to a maximum above the tropopause). The size of this underestimation depends on the size of the region of non- r^2 scatter, which is variable (as we have seen earlier). An ideal solution would be to evaluate Q_0 in the region where aspect sensitivity decreases again, above about 18 km (as reported by *Hocking et al.* [1990], and seen in our observations of Chapter 4). However, our lack of profiler data above this altitude—measurements are only taken up to 19.5 km—prevents us from making a detailed comparison. An examination of the few cases where the region of low aspect sensitivity is somewhat larger indicate that the correction is in the correct sense, and is about the correct order of magnitude, to bring the radar and sonde tropopauses into better agreement.

5.4 Real-Time Implementation

Given the nature of the atmosphere over Adelaide—the climatological double tropopause feature described above, as well as frequent tropopause folding events due to frontal passages—it is often desirable to look for more than one tropopause level. We outline an algorithm developed to perform multiple scans for significant levels and, in the next section, a complementary method of following such multiple layers using cluster averages in time.

Having constructed the $Q(r)$ -profile from returned power in the manner described previously, the immediate task is to identify a level which satisfies our model as a tropopause height. We will comment later on the applicability of the model; for now, let us call any altitude which fits the model assumptions a tropopause height. That is, our aim is to find an altitude r_{trop} that satisfies Equation 5.11.

The first task is to find a suitable value of Q_0 . As demonstrated above, $Q(r)$ tends to peak around the level of the tropopause; in fact, early tropopause identification routines simply took the level of the enhancement to be the tropopause. This resulted in an overestimation of the tropopause height, generally by a few hundred metres.

The tropopausal enhancement is due to the increase in stability, and the concurrent increase in the generalised gradient of radio refractive index (M). If there is a split tropopause, then multiple enhancements over the background exponential decrease in returned power would be expected. However, it is the presence of enhancements which first allow us to identify a tropopause level, and the location of enhancements in an objective manner is our first step.

Starting at the lowest radar altitude, the algorithm searches for a local maximum, the degree of which can be set as desired; three range gates either side of a level is a good compromise between detectability and statistical significance. This sets a resolution limit on the detection of layers which are close together in altitude, but such layers will be apparent over longer periods than a single retrieval, and hence detected by the cluster averaging technique discussed later. It is certainly possible that only one of two nearby layers would be detected, but the degree of faith one would have in layers with less than three range gates separation is questionable.

With an enhancement selected, the $Q(r)$ profile above that level is averaged for some distance, typically of the order of ten range gates; we label the average as Q_0 . If the chosen enhancement corresponds to the tropopause, then this average will be in a region where the model assumption ($dT/dz \sim 0$ where $Q(r) = Q_0$) holds, and we can linearly interpolate below the peak enhancement to find a range satisfying Equation 5.9. Scanning of the range-corrected power profile continues, enabling multiple layers to be identified. If the model assumptions are not met for a given enhancement level, there will be a breakdown at some stage of the evaluation. For example, an enhancement due to random fluctuations will not have a region where $Q(r)$ is near-constant aloft, biasing the Q_0 guess and resulting in an inability to fit a value of $Q_0 - 0.46$ to the $Q(r)$ profile below the enhancement. The possibility of misidentifications due to statistical variation always exists, of course, and we deal with that case in the next section.

5.5 Sources of Estimation Error

We have seen in Figure 5.2 that the sonde-determined tropopause height is almost exclusively located just below the maximum in $Q(r)$ (and, by extension, $\mathcal{P}(r)$; see Figures 4.16 and 4.18, Figure 6.19, and Figure 7.7, for example). The very few cases where the main tropopause approaches or exceeds the level of the returned power maximum are almost certainly statistical fluctuations (although see later for a discussion of the situation when the tropopause is marked by a single sharp change in temperature gradient). The upper tropopause levels are more indistinct, but are certainly associated with (albeit smaller) maxima. The tropopause detection algorithm fails to locate these smaller maxima because they do not pass the significance criterion; the sonde-determined upper tropopauses in panels (9) and (11)–(13) are not associated with radar estimates, for example. Only when the upper enhancement is strong (panel (1), for example) will it pass the significance threshold of being well-separated from the background.

We now consider a detailed analysis of the possible sources of error in the profiler-derived tropopause height estimates, starting with the observed offset between radar and sonde values. There are a number of possible explanations for the small (less than 500 m) but consistent offset between the radar- and sonde-determined tropopause heights. The consistency of the difference tends to invalidate arguments based on statistical fluctuations (which would be just as likely to result in the radar height being above the sonde height, a situation which is rarely seen). Similarly, “moving layer” arguments are unlikely to be convincing, since the tropopause moves both up and down, so any lag effect should work both ways; in addition, the later flights in IOP IV/3 (where the tropopause has returned to near equilibrium) continue to display an offset. Consistent biases in the linear interpolation routine are a possibility, given that the shape of the tropopausal enhancement is occasionally Gaussian rather than triangular or step-like, but the typical examples shown in Figure 5.2 indicate that the offset is present even in cases of strong linearity. Note that the gradient in $Q(r)$ is generally quite shallow in the sub-tropopause region, so that the algorithm is somewhat insensitive to errors in Q_0 (Figure 5.2(8) and others, for example). This is

of considerable benefit in terms of minimising the effects of errors associated with the contributing estimations (such as the calculation of Q_0).

We have assumed that the radar range gates are accurate measures of altitude; in fact, any comparison of radar-determined tropopause heights is utterly dependent on a proper range calibration. Any error in the range markers will be reflected as a systematic error in estimated tropopause heights. Unfortunately, such a calibration is very difficult to implement: unknowns like internal system propagation times are all too often misinterpreted in terms of range. As shown in Chapter 4, a direct comparison of returned power and estimates from sonde thermodynamics can often be used to correct range calibrations, and this method (which is perhaps the only practical direct comparison) provides for adjustments of the order of half a range gate—maybe a couple of hundred metres or so. That this correction is of the same order as the difference between balloon and radar tropopause estimates cannot be taken as evidence of causality, of course; however, the observation of a systematic, near-constant offset over long periods of time (and hence during variable conditions) is certainly suggestive of a range marker problem. The comparative profiles shown in Figure 4.1 are inconclusive as far as assigning a range-marker bias is concerned, as are those in Figure 4.7, indicating that any range marker problem at Buckland Park at the time of the observations was minor. Certainly, any bias visible in those figures appears to be in the correct sense as far as assigning a too-low estimate of tropopause height is concerned, but it should be noted that the underestimation problem in the lower stratosphere will accentuate the *appearance* of an underestimated-in-altitude peak.

The assumption of $dT/dz \sim 0$ for some distance above the tropopause is a good one for Adelaide, in general. However, as we have seen earlier (Figure 5.1) this may not be as good an assumption at sites where the temperature gradient at the tropopause is sharp. In these cases, Q_0 will be overestimated, with a resultant overestimation in the tropopause height. Note, though, that any aspect sensitivity effects (as seen in the range-dependence exponent) will tend to counter this tendency. This effect should be considered in any implementation of the tropopause detection algorithm at such sites.

Finally, we should always bear in mind that an offset could well be due to an actual difference in what the two systems (radar and sonde) are measuring. This could be

due to two reasons: the definition of the radar tropopause not matching the actual application of the sonde's temperature gradient-based algorithm; or a more fundamental difference between the Fresnel scattering model and the realities of atmospheric scatter. As far as the first point is concerned, any further modification to the tropopause detection algorithm (such as an arbitrary change to the offset value in Equation 5.11) would simply be "fiddling", without due scientific reason. With regards to the second point, *Dalaudier et al.* [1994] (following *Dalaudier et al.* [1989]) recently presented evidence of very strong positive temperature gradients within thin layers (typically with a temperature increase of 0.5 K over 10 m) from high-resolution balloon flights. The presence of such sheets is hypothesised by the Fresnel scatter model (amongst others), and *Dalaudier et al.* [1994] show that their contribution to the VHF echo strength is significant, if not dominant. For our purposes, we note that the observed sheets tend to occur in groups in regions of strong static stability, especially just above the tropopause. Hence, the strongest contribution to echo strength will come from these regions. This is exactly what we see in observations of returned power and signal-to-noise ratio (as shown earlier). However, the inverse argument also holds true: in the sub-tropopause region, where these sheets are infrequent, the scatter will *not* be pure Fresnel scatter, and the modelled $Q(r)$ profile will likely be overestimated. This would lead to an underestimate in the radar-derived tropopause height.

While all these effects are likely to contribute in some way, we believe that the major reason for the offset is in the underestimation of Q_0 due to the aspect sensitivity variation (and related range-dependence exponent variation) discussed in Section 5.3. The assumption that $Q(r)$ is near-constant for a few kilometres above the tropopause is perhaps the weakest of all the model's propositions in terms of approximating reality, and the choice of Q_0 is the most sensitive part of the algorithm. Unfortunately, it is also the most difficult factor in terms of applying a correction, for reasons of atmospheric variability outlined earlier with respect to the range-dependence of echoes. An aspect sensitivity or specularity correction is possible from spaced antenna records (see *Briggs* [1992], for example), or from a combination of off-vertical and vertical Doppler beams (which we do not have access to; see *Hocking et al.* [1986] and *Hocking et al.* [1990] for a discussion), and a combination of vertical Doppler with spaced antenna records would

allow for an attempt at resolving the issue; this is the subject of ongoing investigations.

The above discussion deals with the case of determining the precise “radar tropopause height” from a given enhancement. How sure are we that any enhancement actually corresponds to a potential tropopause level? The model we have employed for tropopause identification is dry, so enhancements in Q (due to enhancements in returned power) are assumed to come from temperature-derived perturbations alone: that is, changes in dT/dz . However, in the real atmosphere, changes in moisture content can also produce changes in M^2 , and hence returned power (see Equation 2.16); in fact, the moisture component is dominant in the lower troposphere, as displayed in Figure 4.1). Examination of typical M^2 profiles at a finer scale than shown there reveals that the magnitude of fluctuations in the wet term is comparable with those in the dry term. Hence, it is possible that an enhancement detected in a profile of $Q(r)$ is due to a change in moisture content and humidity gradient rather than temperature gradient; it is similarly possible that such an enhancement would pass the simple algorithmic test of Equation 5.9, and be classified as a tropopause. However, such a level will almost certainly occur in the lowest few kilometres of the atmosphere, generally below 6 km, well away from the unperturbed tropopause. Only in cases of deep tropopause folding is there likely to be significant confusion; in those cases, the continuity of the tropopause level (as followed from the cluster analysis described in the next section, for example) enables one to keep track of the level of interest.

Another source of “false” enhancements could be random (statistical) fluctuations in the measured returned power. The standard deviation associated with each estimate of $Q(r)$ is about 0.5 dB (from the spread of values at each range gate in Figure 5.2). However, a random enhancement is highly unlikely to pass the “local maximum for three range gates either side” rejection criterion, especially when individual profiles are averaged. Rarer still is the random enhancement which then also satisfies Equation 5.11.

Given this comprehensive discussion of possible sources of error, the agreement between the sonde and profiler tropopause estimates seen in Figure 5.2 is remarkably good. In the few cases where estimates differ widely, the explanation is almost always

obvious upon examination: enhanced moisture layers are a good example of this. Statistical fluctuations are somewhat harder to identify from individual profiles, but the use of outlier-rejection averaging techniques serves to eliminate them from consideration. We shall examine such a technique, designed specifically for the study of multiple moving layers, in the next section.

5.6 Cluster Time-Averaging

Once a set of tropopause estimates has been constructed from a number of observations over time, some sort of averaging is an aid to interpretation. However, because each power profile returns up to three tropopause height estimates—that is, up to three independent estimates of altitudes which satisfy the algorithm—a simple numerical average is insufficient to describe the state of the atmosphere. The concept of clustering—finding self-similar groups in a dataset—presents an elegant method of dealing with the problem. Cluster analysis enjoys a wide body of literature, to which the reader is referred for more detail (see, for example, *Romesburg* [1984], *Kaufman & Rousseeuw* [1990] and *Dasarathy* [1991]). A summary of the applicable concepts, however, is appropriate.

Clustering can perhaps be best introduced by the old adage, “*Birds of a feather flock together*”¹. From a given dataset of n elements (objects), we wish to construct k subsets (clusters) such that the members of a particular cluster are more like one another than they are like the members of any other cluster. There are two major ways of approaching clustering: *hierarchical* and *partitioning* methods.

Hierarchical methods involve processing a set of elements one by one, determining the “best” cluster at each stage; there is a gradual transition between 1 and n clusters. They can be agglomerative, where one sequentially adds objects to clusters (so that $k = n, n-1, \dots, 1$), or divisive, by taking the “worst” fitting object away ($k = 1, 2, \dots, n$). Obviously, the clusters formed during the process are not necessarily very good! The process is computationally efficient, but is open to erroneous decisions “along the way”.

¹As an aside, predictive applications of clustering could be summarised by, “*Judge a man by the company he keeps*”.

Hierarchical cluster analysis applications outnumber non-hierarchical in the literature by more than a factor of ten [Romesburg, 1984], although mainly in qualitative fields: genotype ancestry of species, for example, where the results can be summarised by a decision or relation tree, linking one “generation” to the next.

Our needs, however, are better served by a more quantitative approach. Basic partitioning methods involve the construction of k clusters such that each object belongs to exactly one cluster. We try to find a “good” partition, where objects in the same cluster are “close”, or related in some way, while objects in different clusters are distinct. The idea of a distance scale, or *metric*, is central, since it forms the basis of grouping decisions. In our (one-dimensional) case, the metric is simply the absolute arithmetic difference between a tropopause height estimate and the mean of the associated cluster ($|h_i - h_0|$); a least-squares method would be an alternative, although the effect is the same. Mathematically, we wish to minimise the total metric within a cluster, while maximising the between-cluster metric.

The simplest clustering method would be to choose k “representative” objects, and assign the remaining objects to the nearest representative. The initially selected representative objects obviously need to be central to the clusters; an obvious improvement to the method would be to allow the clustering to occur around a point which is not necessarily an object. This is known as “partitioning around medoids”, and minimising the separation (or, more generally, the average dissimilarity) of the objects to others in the same cluster is a straightforward technique.

Our method is based on the variance minimisation technique of *Forgy* [1965], which proceeds as follows. An initial partition of k clusters is constructed through the use of arbitrary seed points. Each of the n objects is then assigned to the cluster with the nearest seed, and the centroid of each cluster is calculated; this is the new seed point. Object assignments are changed, varying the cluster centroids, until no further change occurs—the optimum state has then been found. *Jancey* [1966] extended the method, varying the old seed points by linear combination with the centroids to create the new seeds; *MacQueen* [1967] introduced the k -means method, using both old and new centroids on each iteration so as to more rapidly approach the optimum cluster assignment.

The main problem with variance minimisation techniques comes from outliers, which must be dealt with on a case-by-case basis. A secondary problem arises if clusters are close together, since natural variability—in our case, geophysical and statistical “noise”—will make absolute segregation into distinct clusters difficult. This latter problem is generally intractable as far as our applications are concerned: as the data series increases in length, statistical variability is reduced but geophysical variations become more important. That is, it is practically impossible to discriminate between two close layers in the short term because of statistical fluctuations, and in the long term due to layer motion and evolution.

A technique known as “fuzzy” analysis allows for an extra degree of decision-making, in a manner which can be implemented in software. Essentially, objects are allowed to belong to more than one cluster on a “time-sharing” or likelihood basis. This blurring of cluster boundaries gives the method its name, and allows one to categorise an object by saying, for example, “Object n_0 belongs to cluster $k = 1$ at 90%, cluster $k = 2$ at 7%, and cluster $k = 3$ at 3%”. Objects which do not pass some significance threshold can be classed as outliers, although care must be taken to ensure that failure to achieve significance is not due to a poor choice of the number of clusters (consider the case of clustering a distribution consisting of three distinct layers into just two clusters, for example). The output of a fuzzy algorithm is an $n \times k$ matrix of cluster membership probabilities, although this completeness presents its own problems as n and/or k become large: interpretation becomes difficult, and computation time increases rapidly. In practise, one usually falls back on “hard” decisions, by assigning an object to the cluster with which it has the largest membership coefficient.

Our method assimilates the most appropriate parts of the above techniques to produce a tailored algorithm for average tropopause height retrieval. Data points are initially grouped into three clusters with limits of 8 km and 10 km (*i.e.* the clusters contain estimates of less than 8 km, between 8 and 10 km, and greater than 10 km); any empty cluster is assigned an “invisible” seed point (8 km, 9 km, and 10 km respectively), which is not used in cluster calculations. We then calculate the mean of each cluster, and evaluate the average of the two highest means, and of the two lowest means—these two numbers become the new limits for the next cluster iteration. Points are rebinned

and the steps above repeated, until no significant change occurs in the cluster averages. By keeping track of the cluster assignments, it is possible to calculate the degree of variance away from the cluster mean for each individual point. The worst outlier in a given cluster can be eliminated, and the cluster mean and variance recalculated. If there is a significant difference, this point can be discarded as an outlier and the next worst point selected for similar treatment; otherwise, the cluster average is finalised and the next cluster considered. Since we have already determined that the initial clustering produces the best clusters (in terms of data point spread), and lacking any other way of discriminating between erroneous estimates and the state of the real atmosphere, it is not worth considering the discarded points in other clusters.

The number of points in a cluster give a good first-guess estimate of the reliability of a particular cluster average. For typical conditions, with twelve returned power profile estimates in an hour and a three-pass tropopause determination scan, there are 36 possible estimates. Generally, there are never more than twelve estimates of any given atmospheric layer, since only rarely will a single layer return more than one estimate from an individual profile. In practice, though, first-guess clusters *can* have more than twelve estimates due to spurious points; many are eliminated by the outlier rejection scheme, but (as mentioned earlier) there is no way to identify as “bad” a data point which is close in value to a real layer. Tighter rejection criteria (*i.e.* a lower permissible cluster variance before acceptance) obviously improve confidence in layer identification, but also decrease the number of useful estimates, especially in an evolving atmosphere with moving layers.

5.7 Case Studies

The individual profiles of Figure 5.2 are convincing evidence of the general applicability of the tropopause detection algorithm, at least in broad terms. We now examine the behaviour of the technique over time, and the performance of the cluster averaging. We have chosen a number of datasets on the basis of interesting features, rather than the absolute accuracy of the radar/sonde comparison. There are a large number of cases where a single, stable tropopause exists for a couple of days at a time, and the radar

returns similar consistent estimates (albeit just under the sonde values). However, it is more valuable to examine the cases where such excellent performance does not occur, in an effort to identify possible sources of error and subsequent improvement to the algorithm.

Figure 5.3 displays data from a late-summer IOP during SAFEx (see Appendix C for details). At this time, the radar was collecting vertical Doppler data to a height of just 16.5 km, so it is not surprising that the upper tropopause (near 15–16 km) is not detected: the enhancement is too close to the upper limit of the dataset to pass the “relative maximum in Q ” criterion. However, note how the radar detects the lower tropopause early in the dataset: this layer is detected by the first sonde flight (about four hours into the IOP), but the radar retrievals show the persistence of the layer for some hours either side of the sonde flight. The middle tropopause (near 11 km) which dominates the radar returns from 15 hours onwards is also confirmed by two sonde retrievals (at 19 and 29 hours). However, the radar tracking indicates that this layer exists across nearly the whole dataset, apart from some confusion near the “break” region around 10 hours into the IOP. Note how the three-cluster approach picks up a triple-layered structure around 11 hours, which is visible in the raw data as well as the averages. The region around 9 hours is confused due to lack of profiler data, but the cluster average shows indications of layer motions which the sonde flights cannot detect due to their relatively poor temporal resolution.

A longer dataset is displayed in Figure 5.4, which is taken from the major case study we discuss in Chapter 6. The deep tropopause fold around 1200 UTC 29 August is clearly detected by the profiler retrievals, while the movement of the upper tropopause (near 15 km) matches well with the radiosonde estimates. Note how the raw profiler estimates of the lower tropopause near 0600 UTC 30 August are tightly clustered: this indicates that the layer enhancement is quite sharp and persistent in time. This feature of the radar observations can be quite helpful in interpreting the nature of stable layers: “diffuse” layers, with a relatively wide spread of profiler estimates, can be compared with stronger layers, unlike radiosonde estimates where a single point is returned.

There is a major difference between the techniques late in the IOP, around 0000 UTC

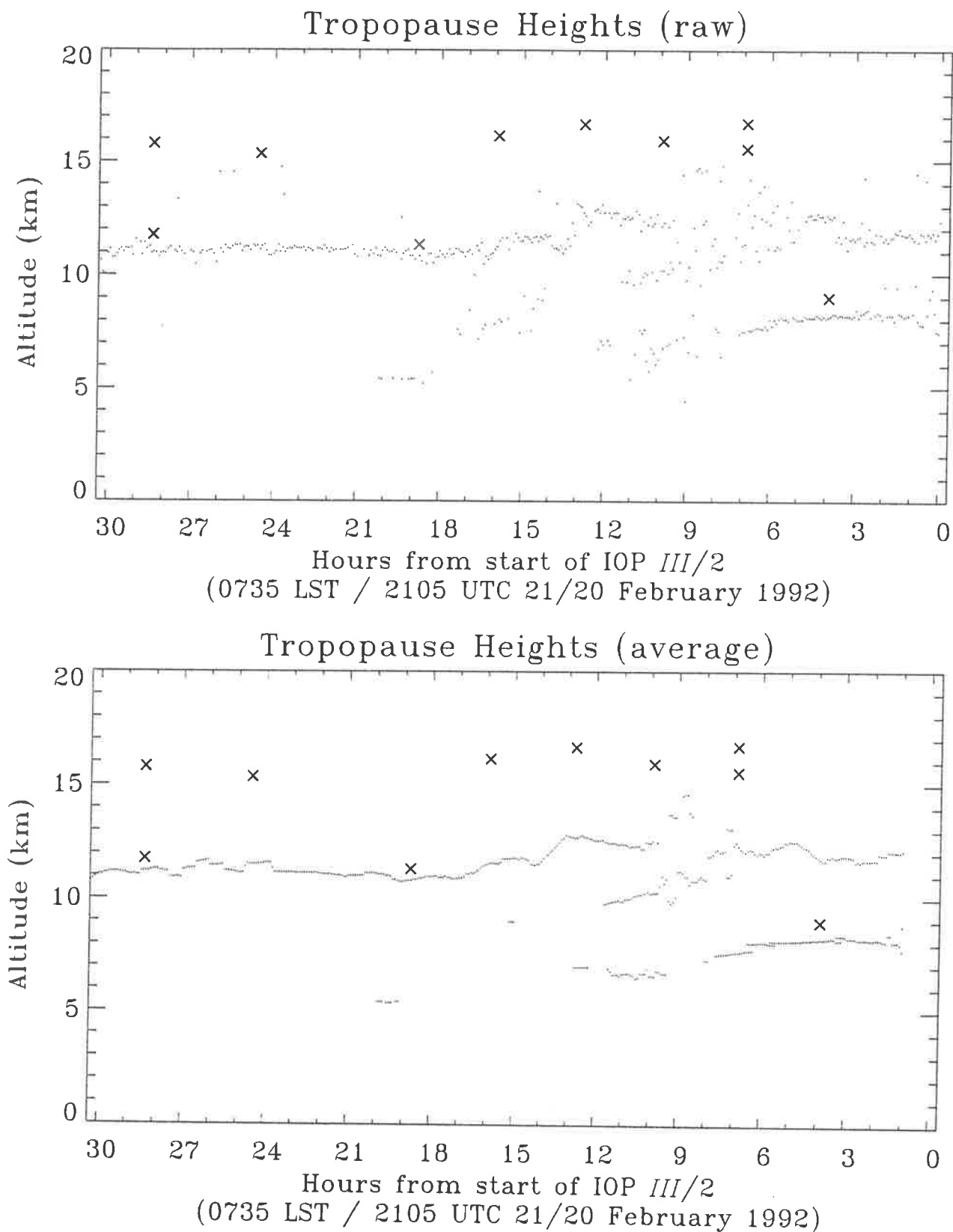


Figure 5.3: Radar-determined tropopause heights for SAFEx IOP III/2: the upper panel shows the raw, five-minute estimates, while the lower panel is a three-cluster mean based on an hour of data. The sonde estimates are denoted by $\times$'s. Time runs from right to left.

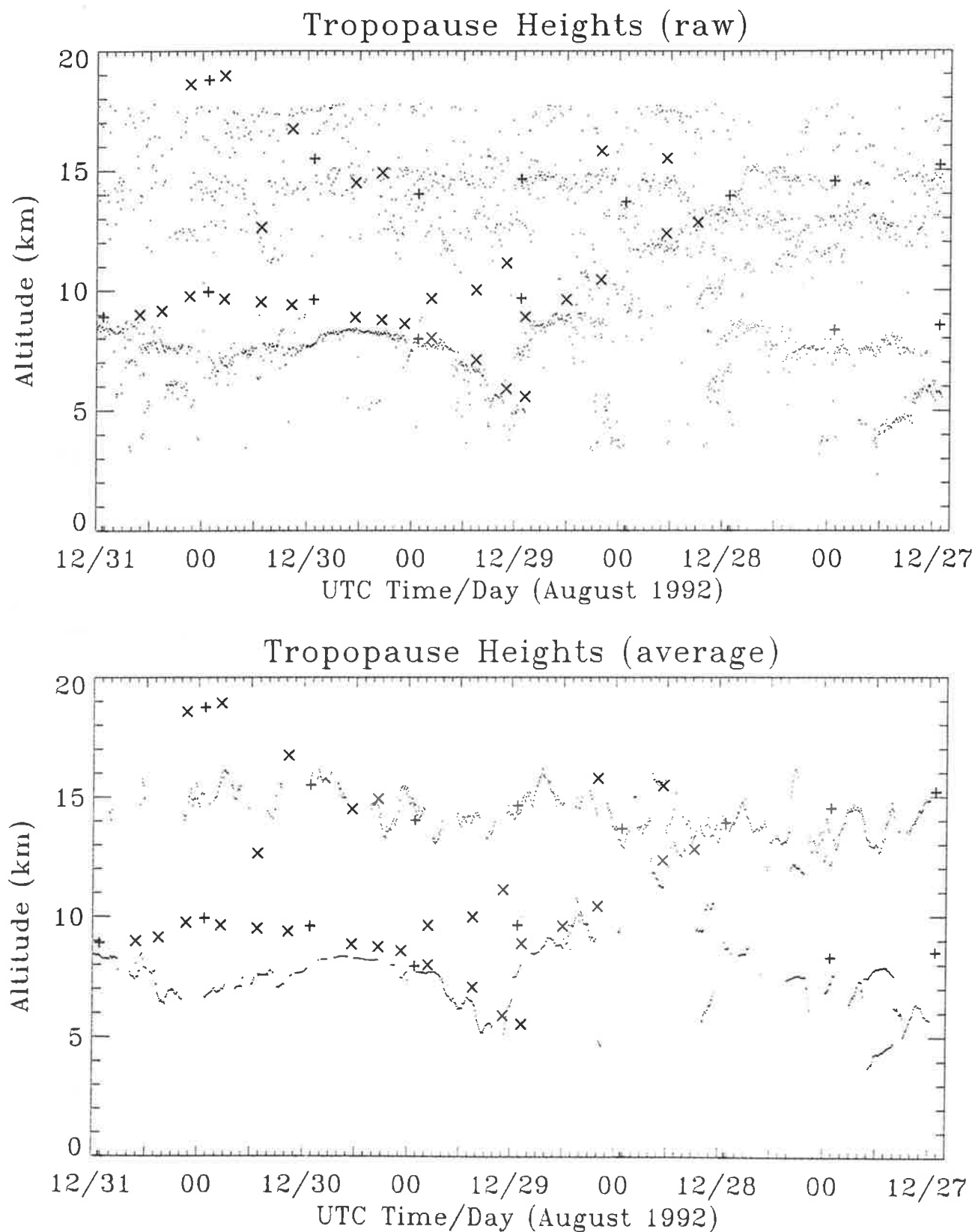


Figure 5.4: As in Figure 5.3, but for SAFEx IOP IV/3. Additional sonde estimates from Adelaide Airport are denoted by +’s.

31 August, where the profiler consistently detects a layer some two to three kilometres below the sonde estimate. A number of fields displayed in Chapter 6 help with interpretation in this case. For example, the profiler signal-to-noise ratio (Figure 6.19) clearly shows the base of the enhancement which the tropopause detection algorithm is following; an enhancement in the Brunt-Väisälä frequency (with a magnitude between typical tropospheric and stratospheric values) is also visible, in Figure 6.18. Examination of the potential temperature cross section (Figure 6.10) reveals a more stable region beneath the sonde-analysed tropopause in this region (the 312 K potential isotherm for 24 hours from 1200 UTC 30 August). Even with the range and pressure normalisations, the profiler tends to see enhancements which are *lower* in the atmosphere as relatively *stronger*; in this case, the effects are obvious. Although the layer seen by the profiler does not completely and fully satisfy the meteorological definition of a tropopause, it is close enough (under the assumptions of the detection algorithm) to be classified as such. A better range resolution would help in situations such as this, since the multiple layers which are barely visible in the signal-to-noise ratio around this time (in Figure 6.19) would be individually identified, rather than being interpreted as a single, broad layer.

Figure 5.5 displays a case where both upper and lower tropopauses are detected strongly and continuously over the duration of the IOP. However, the raw data displays three clear signals (with traces of a fourth!) between three and twelve hours into the dataset. The two upper levels are too close together for the clustering algorithm to discriminate between them, and it analyses them as a single, undulating layer. The lower two layers are somewhat different: the very bottom signal is far stronger, and is clustered by itself (barring a short period around five hours into the IOP). Again, the vertical resolution of the profiler is an important factor; on occasions, an investigation of the raw data (rather than simply the cluster average) is appropriate, especially when the assumptions of the cluster averaging, in terms of the number and closeness of layers, are poorly satisfied.

Finally, and to indicate the more general applicability of our results, we have applied the tropopause detection and cluster mean algorithms to the Darwin profiler data discussed in Chapter 4. The results for the month of January 1992 are displayed

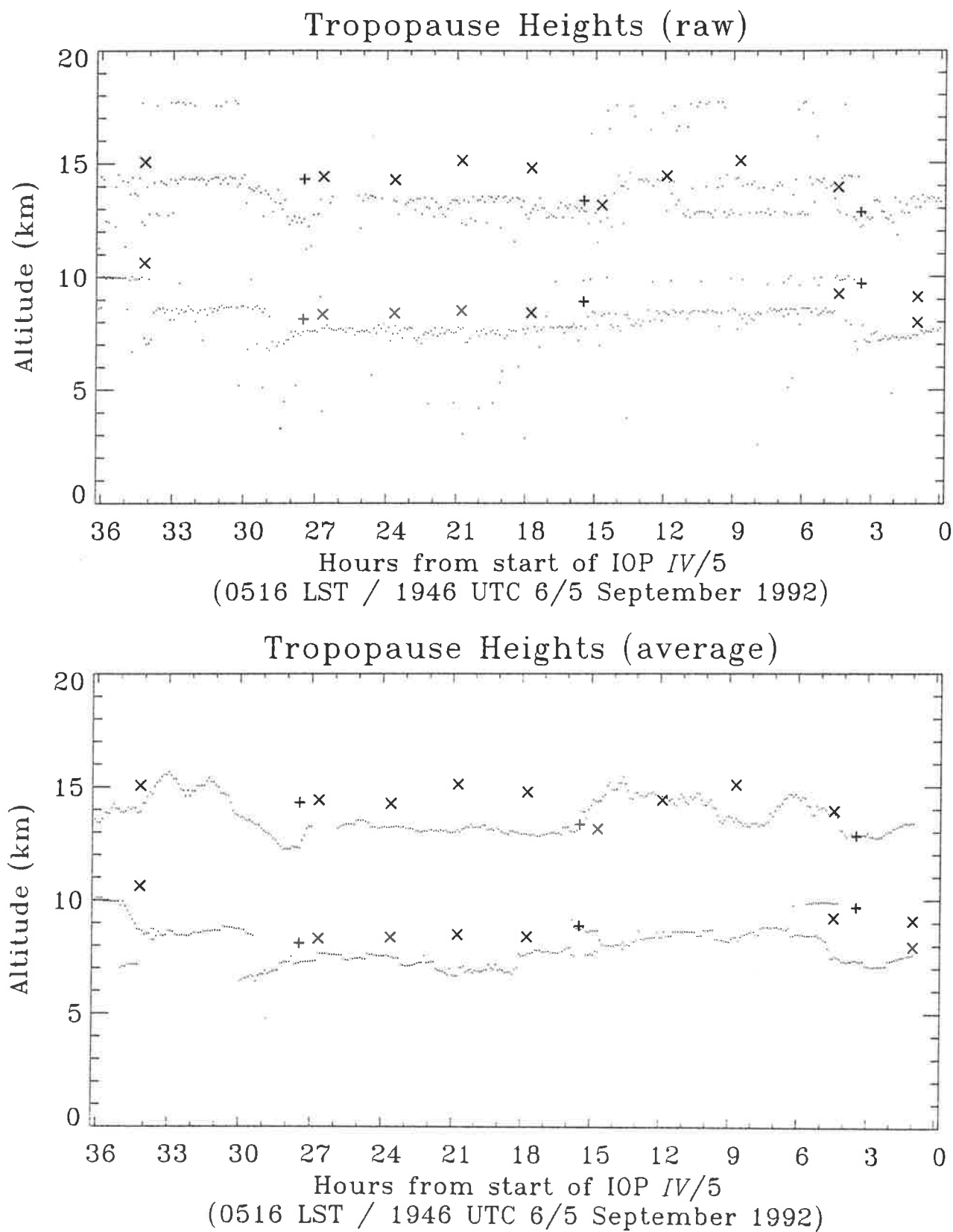


Figure 5.5: As in Figure 5.4, but for SAFEx IOP IV/5.

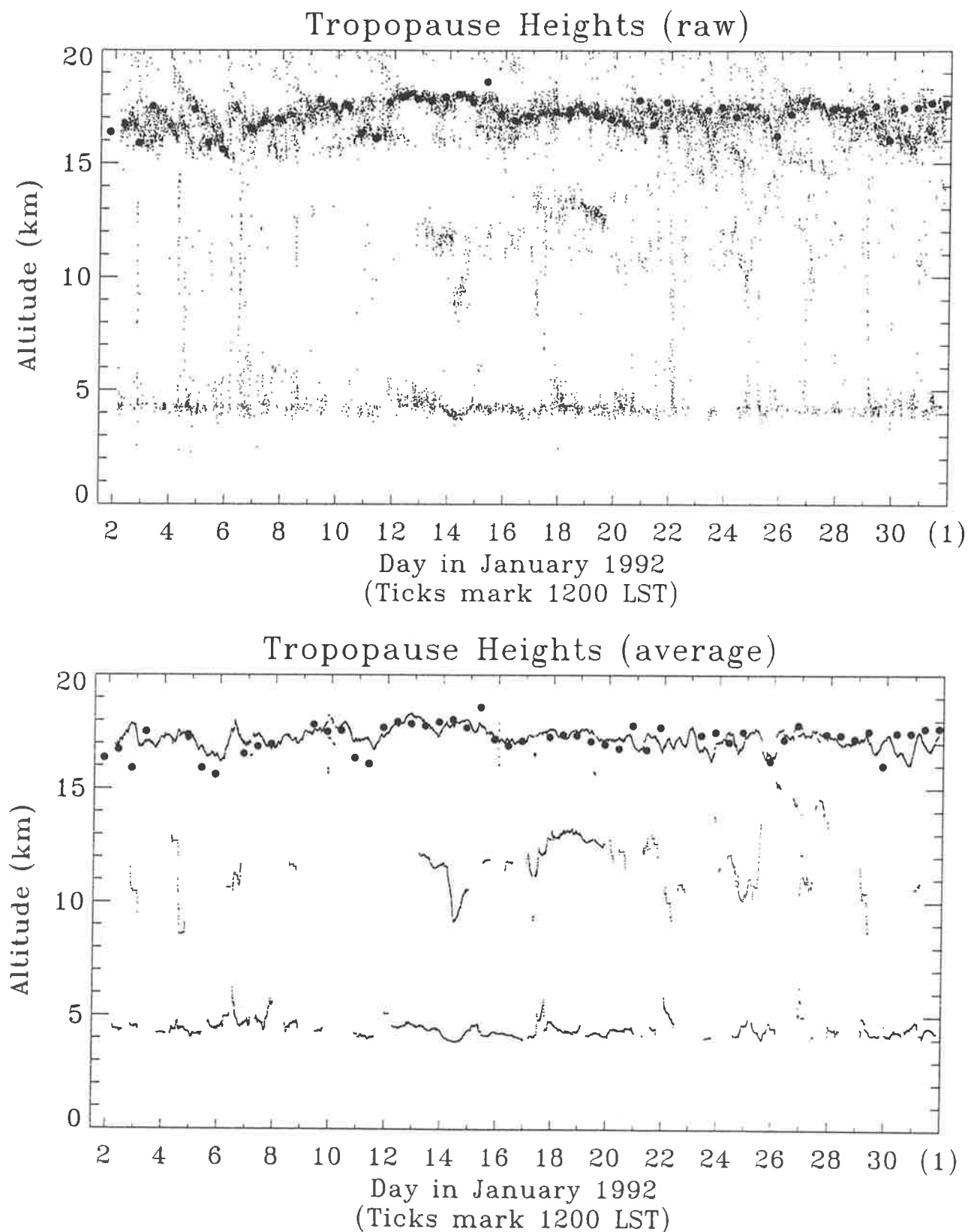


Figure 5.6: Radar-determined tropopause heights for Darwin, January 1992: the upper panel shows the raw four-minute estimates, while the lower panel is a three-cluster mean based on six hours of data. Sonde estimations from the twice-daily flights are indicated by solid circles. Note that the time axis is in *days*, and runs from left to right.

in Figure 5.6, and the tropopause around 17 km is detected quite accurately. In particular, the *changes* in tropopause height, the short-term rises and falls, are reproduced well in the raw data if not in the cluster mean. Compare the raw and averaged data on January 5, 11, 26 and 30 as good examples of this. Note how the sonde tropopause identifications overlay the raw profiler estimates throughout the month; the finescale variations in the cluster averages can be misleading, in the same way as running means (to which the clustering approach is related). The sharp tropopause at Darwin (Figure 5.1) is likely to be associated with a different distribution of scatterers above that level; as we have discussed above, the Q_0 underestimation is then less significant, resulting in a better tropopause estimate.

The enhancement near 4.5 km can be seen in the profiler observations (Figure 4.19), and has been identified in Section 4.5.3 as due to the radar brightband at the freezing level. The sporadic enhancements around 11 km are due to features in the upper troposphere (such as those between January 18 and 22), and spurious retrievals (the vertical line-like features apparent in the raw data, for example).

We believe that these types of diagrams, together with the related discussion, is far more appropriate than displaying a collated scatter plot of radar versus balloon tropopause heights. The effects of local conditions can dominate the radar estimates in a manner for which the reasons are readily apparent from examination of the time series, but which are hidden in a grouped plot. In addition, the sonde retrievals are usually limited to two estimates per profile, while the radar can detect more (as we have shown above); objectively deciding the appropriate level to compare can be difficult.

5.8 Summary and Extensions

We have demonstrated that the tropopause detection algorithm as applied to vertical-beam profiler data can identify stable layers which satisfy a version of the WMO definition of the tropopause. However, one must be careful in attempting to draw a direct link between definitions of the tropopause based on temperature gradient, and layers as detected by the profiler. In many ways, this is similar to the conceptual differences between the thermal tropopause and the dynamical tropopause

(as defined by potential vorticity; see Chapter 6 for an application of the dynamical tropopause), although not to the same extent. The “radar tropopause” is located just below an enhancement in returned power (or signal-to-noise ratio), and misidentifications can occur when enhancements due to other effects (such as moisture) appear in the returned power profile. The current implementation of the algorithm appears to underestimate the midlatitude tropopause by a couple of hundred metres, although the tropical tropopause is identified nearly perfectly. A likely explanation involves underestimation of the Q_0 parameter due to aspect sensitivity effects in the lower stratosphere. Correcting for the changing nature of the scatterers would be an obvious extension, aiming for accurate hourly estimates of tropopause height. This is far better than the current (sonde) resolution of twice-daily estimates, from which it is difficult to make predictive forecasts. Hourly or better estimates allow for feature tracking on the mesoscale (or even smaller scales) which can be of help in identifying rapidly evolving situations.

The retrieval of atmospheric parameters is the goal of any experiment in atmospheric physics. If we can relate these parameters to other, perhaps more directly useful quantities, then so much the better. One example of this is the use of radar data to retrieve a temperature profile of the atmosphere from a given profiler sounding, as originally suggested by *Gage & Green* [1982b].

The method is essentially as follows. From Equations 5.2 and 5.3 we have :

$$\mathcal{R}(r) = \beta \left(\frac{p}{T^2} \left[\frac{dT}{dz} + \Gamma_d \right] \right)^2 \quad (5.12)$$

Hence, we can solve for the temperature gradient :

$$\frac{dT}{dz} = \frac{T^2}{p} \sqrt{\frac{\mathcal{R}(r)}{\beta}} - \Gamma_d \quad (5.13)$$

If we substitute the usual exponential decrease expression for pressure (Equation 5.4) then we have a differential equation which can be solved iteratively for the temperature as a function of height, given an initial guess (isothermal atmosphere, for example) and suitable boundary conditions (such as the height and temperature of the tropopause). Note that we can use the tropopause detection algorithm discussed in this chapter to determine the height of the tropopause, which can be related climatologically to the tropopause temperature. Figure 5.7 displays an example of this,

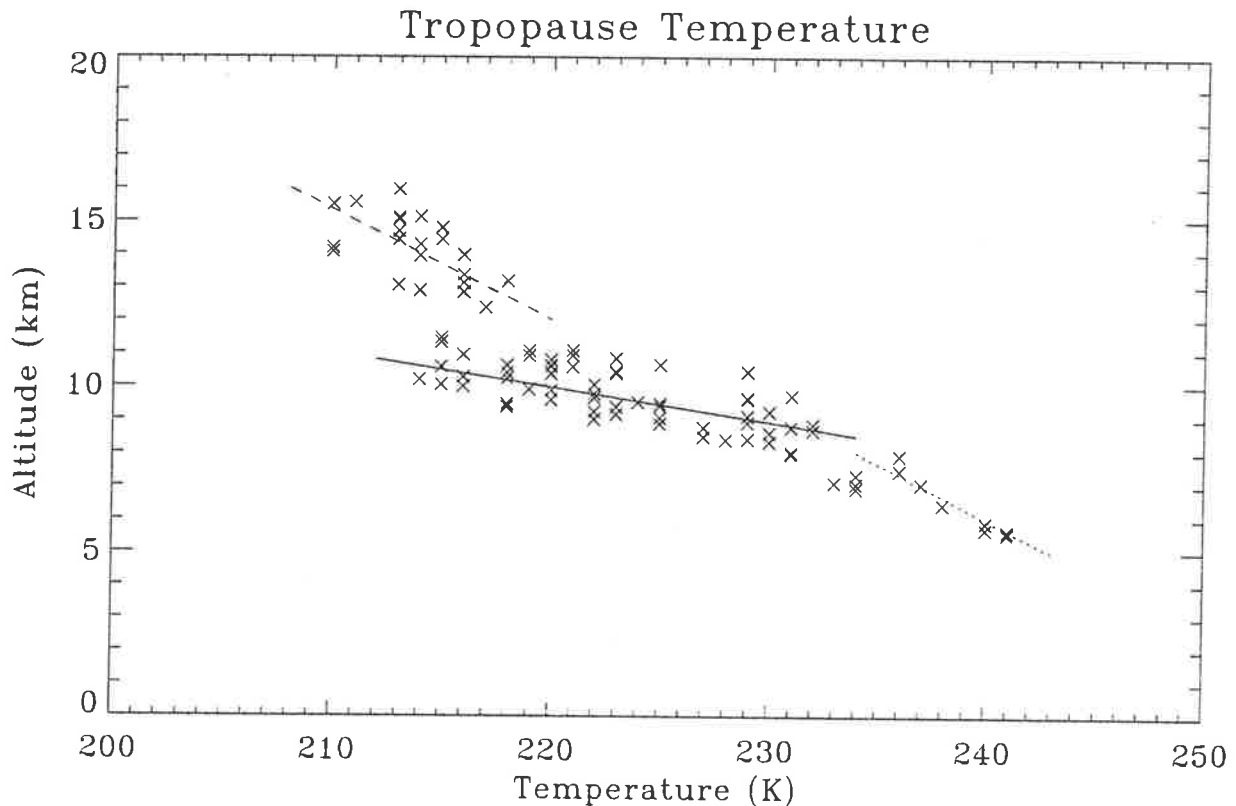


Figure 5.7: Tropopause temperature versus tropopause altitude for SAFEx IOP IV/3. Dotted, solid and dashed lines indicate lower, middle, and upper tropopause distributions respectively.

where three distinct groupings are evident; given the sparse data, we do not concern ourselves with a “best fit” analysis. The grouping occurs because of the characteristic temperature lapse rates of different regions; the upper tropopause, for example, tends to occur at the top of a near-isothermal region, while the lowest tropopause heights are associated with deep tropopause folds.

This method as described above cannot be implemented at present on the Buckland Park array, since it requires the absolute returned power (in the proper physical units) and the radar parameters (included in β) to solve the differential equation. Variations on the method are currently under investigation, with the aim being to eliminate β in an indirect manner, similarly to the actual tropopause detection algorithm.

The discussion so far has assumed a dry atmosphere (barring the consideration of moisture layers being interpreted as temperature-gradient layers). The question now

arises as to the possibility of retrieving *humidity* profiles from the radar data. Our basic idea is to use a climatological dry background of effective returned power (derived from historical records of M_d , as in Chapter 4), and extract the wet component of returned power from the difference between the observed returned power profile and the dry background. As one might imagine, this presents considerable practical problems. Preliminary investigations indicate that the errors involved are large, with the main concerns being confusion between local temperature inversions and moisture-induced layers in the observed power profiles. Certainly, the moisture component is dominant in the lowest few kilometres (see Figure 4.1, for example), but the coarse resolution of the Buckland Park profiler (1 km, oversampled at 500 m intervals) presents severe problems in comparisons. Other concerns include the obvious assumptions regarding climatological behaviour (diurnal variations and intra-seasonal variability), the size of gradient quantities such as dq/dz compared to the magnitude of q , statistical fluctuations in returned power becoming confused with humidity variations, and local temperature inversions producing fluctuations away from the climatological mean. All of these have been encountered in the course of attempted moisture profile extraction, but we remain cautiously optimistic about future developments. In particular, higher resolution observations, the use of UHF systems which are far more sensitive to moisture than VHF profilers, and experiments explicitly designed for this purpose may well provide the keys to the successful extraction of moisture profiles from profiler data.

Chapter 6

AN INTENSE MIDLATITUDE CYCLONE

Research is what I'm doing when I don't know what I'm doing.

— *Werner von Braun*

6.1 Introduction

Late in August 1992, an intense midlatitude cut-off low developed to the south of Australia. The subsequent passage of this system across south-eastern Australia and the Great Australian Bight coincided with the final stages of the South Australian Fronts Experiment (SAFEx), a collaborative observing campaign involving the Centre for Dynamical Meteorology and Oceanography at Monash University, and the Department of Physics and Mathematical Physics at the University of Adelaide. Radiosondes with Omega windfinding capability were launched from the Buckland Park Research Station, and the 54.1 MHz wind profiling radar at Buckland Park was operated continuously throughout the campaign. Combined with automatic weather station records at the field site, regular Bureau of Meteorology measurements, and integrated synoptic analyses, our observations contribute to a dataset of detail unprecedented in the region for the purpose of mesoscale analysis. Full details of the SAFEx programme and data archives are given in Appendix C.

Our aim here is to examine the origin and structure of one particularly notable

event, a cut-off low (intense midlatitude cyclone) recorded as SAFEx IOP IV/3 (27–31 August 1992). The high time resolution (21 sondes in 96 hours, with flights every three hours around the time of passage, and 5-minute profiler retrievals) allows a uniquely in-depth investigation, especially with regards to the mesoscale analysis. The passage of the system initiated major flash flooding in the Adelaide Hills, resulting in the loss of two lives when a caravan park near Cudlee Creek (25 km northeast of Adelaide) became flooded. Strong winds combined with the heavy rainfall to cause over ten million dollars damage to houses, roads and bridges [Furler, 1993]. An initial analysis of this event has been presented by *Mills & Wu* [1995], who concentrated on the synoptic system and its relationship to the onset of severe convection. Our objective with this chapter is to provide further insight into the mesoscale system from a consideration of radiosonde and profiler observations. Additionally, the variety of data sources available allows us to evaluate the performance of the wind profiler in terms of operational short-term forecasting (“nowcasting”).

The characteristics of the Buckland Park profiler have already been discussed in Chapter 3. The observational mode during IOP IV/3 alternated vertical-beam Doppler estimates of vertical velocity with spaced antenna measurements of the horizontal wind field on a five minute cycle, with 500 m resolution between 2 km and 15 km altitude. The Väisälä MARWINsonde system, measuring profiles of pressure, temperature, humidity, and wind every 50 m up to about 20 km, has similarly been described in Section 3.2. Eighteen SAFEx flights have been supplemented in this chapter with three Bureau of Meteorology sonde flights from Adelaide Airport, covering 24 hours prior to the start of the SAFEx flights. We have used the thermodynamic profiles from these flights, which are of functionally identical quality to the SAFEx data; the Adelaide Airport wind data are not used, since they are determined from radar-tracking of the sonde and hence have large inherent errors and a poor vertical resolution. At the surface, events were identified and timed on the basis of one-minute resolution automatic weather station (AWS) observations of pressure, temperature and wind (see Section 3.3). Finally, GMS infrared satellite imagery and 150 km grid-spacing objective analyses with 15 pressure levels in the vertical (the RASP limited area data assimilation system, as described by *Mills & Seaman* [1990] and *Mills & Wu* [1995])

provide a synoptic scale background to the event.

6.2 Background and Context

Case studies of this type have been carried out almost exclusively in the northern hemisphere, and the few examples from the southern hemisphere have not had access to high-resolution data of the type presented here. If we consider the general case of interactions between fronts and “cold-air vortices”, the climatologies of *Carleton & Carpenter* [1990] and *Sinclair & Cong* [1992] show that interactions are by no means rare; however, the majority of observational studies to date have focussed on cyclogenesis within cold airstreams behind *polar* fronts [*Browning*, 1990; *Reed*, 1990].

Studies of extratropical cyclogenesis in the Australian region are not numerous, and are based almost entirely on routine Bureau of Meteorology analyses and satellite imagery. The work of *Downey et al.* [1979] is worthy of note: by incorporating individual radiosonde profiles with the standard analyses, they achieved superior resolution (although not to the same degree as our current work); and the system studied therein has many features in common with our case (in particular, a lowered tropopause over the surface low, a strong upper-level jet to the north, a North-west cloud band, and heavy rainfall). The work of *McRae* [1955], *Ryan & Meadows* [1979] (where the broad scale features are similar to our case), *Holland et al.* [1987] (looking at East-coast cyclones), *Veldon & Mills* [1990] (an extratropical cyclone producing heavy rainfall in summer), *Mills & Russell* [1992] (a system of tropical rather than extratropical origin, again associated with heavy rainfall) and *Bell* [1982] are all typical of investigations in the Australian region.

The latter work [*Bell*, 1982] is of particular interest, since it concerns interactions between subtropical cloud bands and cold-air vortices over Australia (*Thepenier & Cruette* [1981] consider similar interactions over the Atlantic Ocean). Satellite observations suggest that the evolution of such a system depends on the distance between the cold-air vortex and the front [*Zillman & Price*, 1972]. If the component features remain well-separated, then any interaction is weak and they evolve as distinct entities; if the distance between them becomes too small, strong interaction leading to

secondary development on the front is a possibility. We are interested in a more intermediate case, where the front interacts with the cold-air vortex in such a manner that, at least in satellite imagery, the combined system resembles a mature occluded cyclone. *Browning* [1990] refers to this as an “instant occlusion”, and the process has been studied in some detail by *Reed* [1979], *Locatelli et al.* [1982], *Mullen* [1983], *Browning & Hill* [1985] and *McGinngle et al.* [1988], amongst others; the work of *Carleton* [1981] represents the only work specifically on southern hemisphere instant occlusions. We also note the work of *Wright* [1988], who found that the precipitation associated with midlatitude systems was greater if there was an interaction with tropical or subtropical features. This is consistent with the observations presented here.

We will demonstrate that the instant occlusion process is relevant to the case study of SAFEx IOP IV/3. The northern hemisphere case studies of such events (referenced above) show formation of the occlusion in the cold air poleward of a major polar-frontal cloud band; our case documents an instant occlusion formed on the poleward side of a *subtropical* frontal cloud band (and associated subtropical jet stream). We shall refer to *Browning & Hill* [1985] as an example throughout this chapter, drawing attention to the observed similarities and differences in cloud band structure, jet structure, and tropopause deformation. The lack of high-resolution Australian case studies highlights the need for local work, which can contribute to the store of knowledge for forecasting applications and improve our understanding of the initiation of severe weather.

The remainder of the chapter is developed as follows. Section 6.3 gives an overview of the synoptic situation from model analyses and satellite pictures, including a discussion of the frontogenesis through central Australia. Section 6.4 concentrates on the mesoscale aspects of the event, looking in particular at the profiler and sonde data collected during SAFEx, and the convection associated with the event; our conclusions finish the chapter.

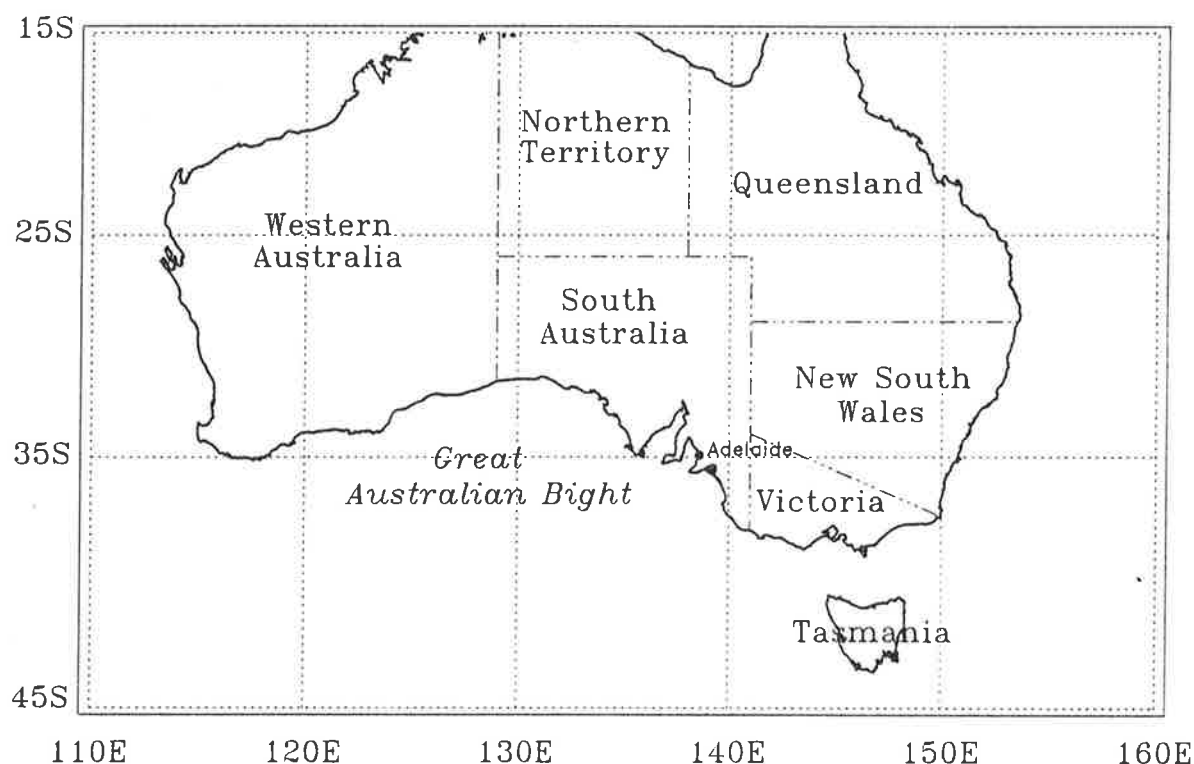


Figure 6.1: A map of the synoptic domain around Australia, with major place names and approximate State borders. A solid dot marks the position of Adelaide (35°S, 138°E).

6.3 Synoptic Analysis

The majority of the analysis in this section is based on the Bureau of Meteorology's regional model (RASP, from Regional Assimilation and Prognosis; see *Mills & Seaman* [1990]). At the time of the experiment, RASP was running on a 150 km resolution grid with 15 pressure levels in the vertical. The reader is also referred to *Bourke et al.* [1995], who use this event as a test case for the Bureau of Meteorology's upgraded Global Assimilation and Prediction system (GASP). The GASP forecasts (at maximum resolution) successfully predict the general situation, although they fail to capture the finer details. Comparison of the RASP and GASP analyses demonstrate consistency, and we are satisfied that RASP describes the synoptic conditions well.

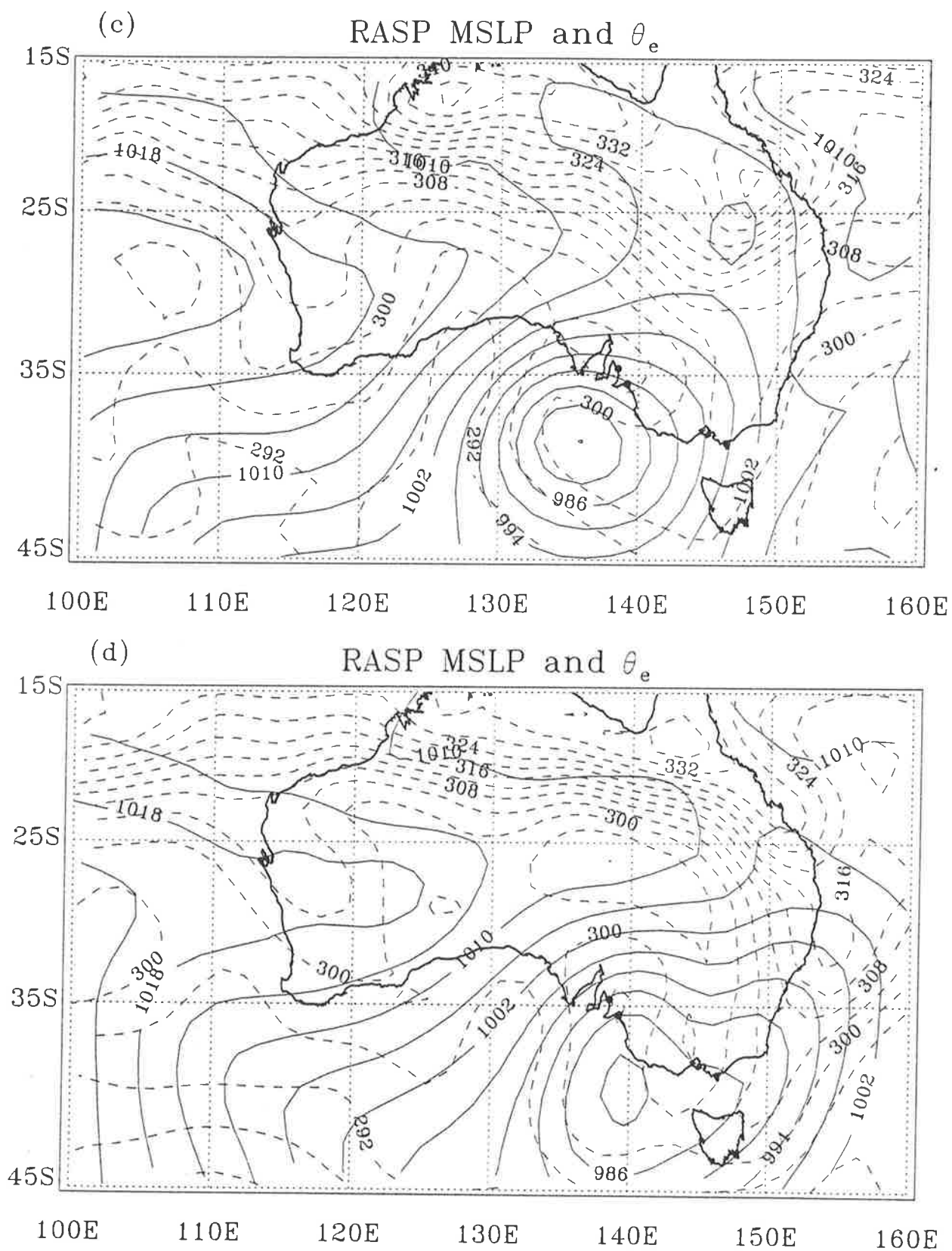
The event examined here developed from a common regional synoptic pattern on 27 August 1992, comprising a cold-frontal cloud band over eastern Victoria and

Tasmania, and a North-west cloud band over Western Australia (see Figure 6.1 for the location of place names mentioned in the text). The front subsequently moved to the east and weakened, while the North-west cloud band strengthened across the continent. Our analysis begins at this point (further details on the origin of the system can be found in *Mills & Wu* [1995]).

Figure 6.2 shows four sequential twelve-hourly analyses of mean sea-level pressure and 900 hPa equivalent potential temperature for the Australian region, starting from 1100 UTC 28 August 1992. The principal feature of interest is a cut-off low pressure system south of Australia, which deepens to a minimum pressure of 978 hPa by 1100 UTC 29 August (Figure 6.2(c)). We note that the deepening rate does not satisfy the “bomb” criterion of *Sanders & Gyakum* [1980], maximising at about 14 hPa over 24 hours¹. The system moves slowly to the east, so that by 2300 UTC 29 August (Figure 6.2(d)) the trough is tilted from the south-west towards the north-east. Strong ridging across central Australia is evident throughout the sequence. Note also the strong gradient in equivalent potential temperature through central Australia: this is the subtropical front, which strengthens with time while the temperature gradients associated with the cut-off low weaken. Figure 6.2(c) clearly shows temperature advection around the low, with warm moist air wrapping to the south-west and cold dry air wrapping around to the north-east. We refer to these features as the warm and cold tongues respectively. Note the small patch of warmer air near the centre of the low in this figure, suggestive of an occlusion. By 2300 UTC 29 August, the southern warm pool has moved over Adelaide and become isolated (Figure 6.2(d)).

The broad features can be seen in Figure 6.3, which shows a sequence of six infrared satellite images taken from *Mills & Wu* [1995]. The North-west cloud band (marked **B**) is associated with the subtropical front [*Tapp & Barrell*, 1984; *Kuhnel*, 1990], and extends southwards across the Australian continent. There are three points to note regarding the North-west cloud band: its southern edge is more distinct than its northern edge; the cloud tops are higher towards the south-east (on the infrared

¹To paraphrase, however, “*All that is gold does not glitter*,” as the eventual development of the system demonstrates.



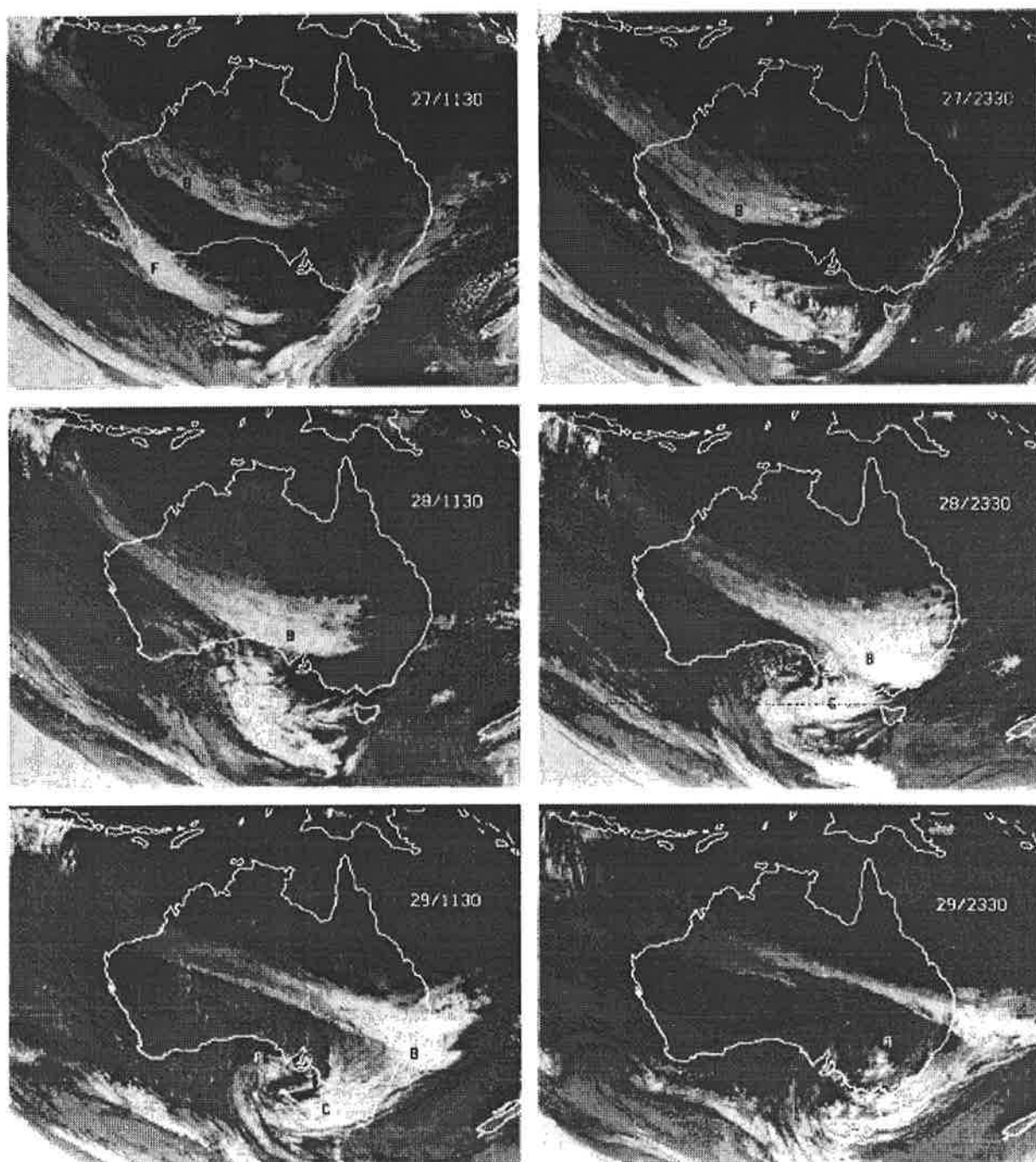


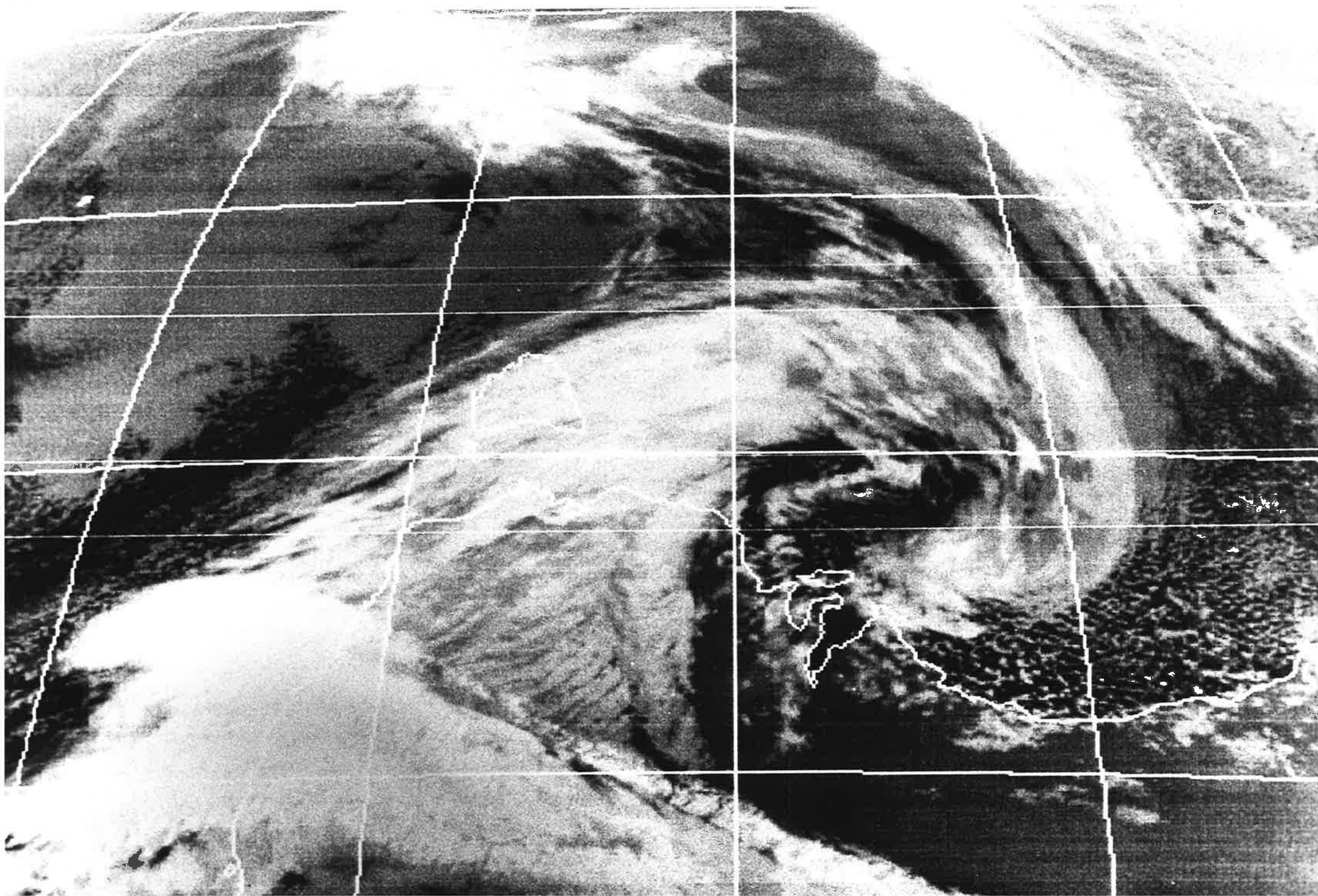
Figure 6.3: Sequence of six, twelve-hourly GMS infrared satellite images, from 1130 UTC 27 August to 2330 UTC 29 August 1992. Letters refer to features mentioned in the text. From *Mills & Wu* [1995]

picture, brighter implies colder, and hence higher); and the cloud curves anticyclonically towards its eastern extent. The cloud band to the south-west of Australia in the initial stages (marked **F**) moves towards the south-east, where it encounters the cut-off low and rotates cyclonically, assuming the shape of an inverted comma (marked **C** in the 2330 UTC 28 August and 1130 UTC 29 August images); this shape is a characteristic of midlatitude cyclones, and in our case is closely associated with the warm tongue. Cloud-top temperatures indicate that the North-west cloud band is composed of upper-level cloud, whereas the comma cloud is located underneath in the mid-troposphere.

Plate I shows an enlargement of the satellite image at 1130 UTC 29 August, concentrating on the area around Adelaide, and is particularly useful in terms of identifying air masses associated with the system. By this time, a cloud-free dry slot associated with the cold tongue has developed to the rear of the warm tongue, while the cloud head (marked with a *white* letter **A** in Figure 6.3) to the north-west of the low has become more pronounced, following the warm air advection around the west of the surface feature. Also note the delta-shaped cloud between the North-west and comma clouds; *Feren* [1995] observed clouds such as these to be a precursor to major cyclogenesis. Of particular interest are the wave-like features in the cloud, which we believe may be associated with gravity-wave activity (see Figures 4.2 and 4.5, and the related discussions). The clearer air visible behind and to the north of the low in the infrared image is cold, dry air descending behind the surface low. A small region of north-south oriented cloud can be seen over the Adelaide area; this area of intense convection was responsible for the flash floods in the Adelaide Hills.

The situation studied here shows a number of similarities to that described by *Browning & Hill* [1985]. Comparison of their Figure 2 with our Plate 1, and making the appropriate “hemisphere inversion”, shows remarkable similarities. The more equatorward cloud (our North-west cloud band) is equivalent to the polar front cloud band (or warm conveyor belt) of *Browning & Hill* [1985]: in both cases it is the highest cloud, in the same relative position, and it turns anticyclonically. The air in this warm conveyor belt originates in the warm region east and equatorwards of the surface low, then ascends and moves polewards of the surface front. *Browning & Hill* [1985]

Plate I (over page): An enlargement of the GMS infrared satellite imagery for 1130 UTC 29 August 1992. Note that the next three numbered pages are deliberately left blank.



also identify a polar trough cloud band, associated with the cold conveyor belt. Its origins are in the warm sector indicated by relatively high wet-bulb temperatures, and it turns cyclonically around the low pressure system. The polar front and polar trough of *Browning & Hill* [1985] are directly analogous to our subtropical front and cut-off low. There is a dry slot in a similar position in both case studies. Perhaps the major difference between the systems is simply that our system is closer to the equator!

Hence, by 1130 UTC 29 August the system has formed an instant occlusion. Although the cloud pattern resembles that of an occluded extra-tropical cyclone, the subtropical front plays the role of the classical cold front, with the North-west cloud band acting as the warm conveyor belt cloud. Similarly, the warm tongue which flows under the North-west cloud band at right-angles is similar to the cold conveyor belt in conceptual models of extratropical cyclones. An explicit trajectory analysis of air parcels which move through the system confirms these relative motions [*Griffiths*, 1996]: trajectories to the north associated with the North-west cloud-band ascend and move southeast to just north of Adelaide, and then turn eastward; more southerly trajectories starting in south-western Western Australia ascend and travel east, then ascend while turning cyclonically around the low.

Figure 6.4 shows the upper-level structure of the system at 1100 UTC 29 August, with contours of potential vorticity (PV) and system-relative wind vectors on the $\theta = 310$ K isentropic surface for the key time of 1100 UTC 29 August. The system speed is calculated from the position of the surface low pressure centre at six-hourly intervals; the pattern of system-relative winds is not sensitive to the system speed. The light shaded area shows where the isentropic surface intersects the tropopause; the dark shaded area is where the surface is located in the stratosphere. In accordance with the suggestion of *Hoskins et al.* [1985], the cut-off low is characterised at upper levels by an isolated cyclonic PV anomaly above the surface feature. Note that weaker cyclonic PV anomalies are aligned with the subtropical front in this figure. Examination of the time-evolution of the PV field (not shown) makes the movement of the upper PV anomaly towards the east readily apparent. *Hoskins et al.* [1985] suggested that if there is low-level warm advection ahead of the upper-level anomaly, there may be positive feedback such that the system can develop. Hence, a region of high PV

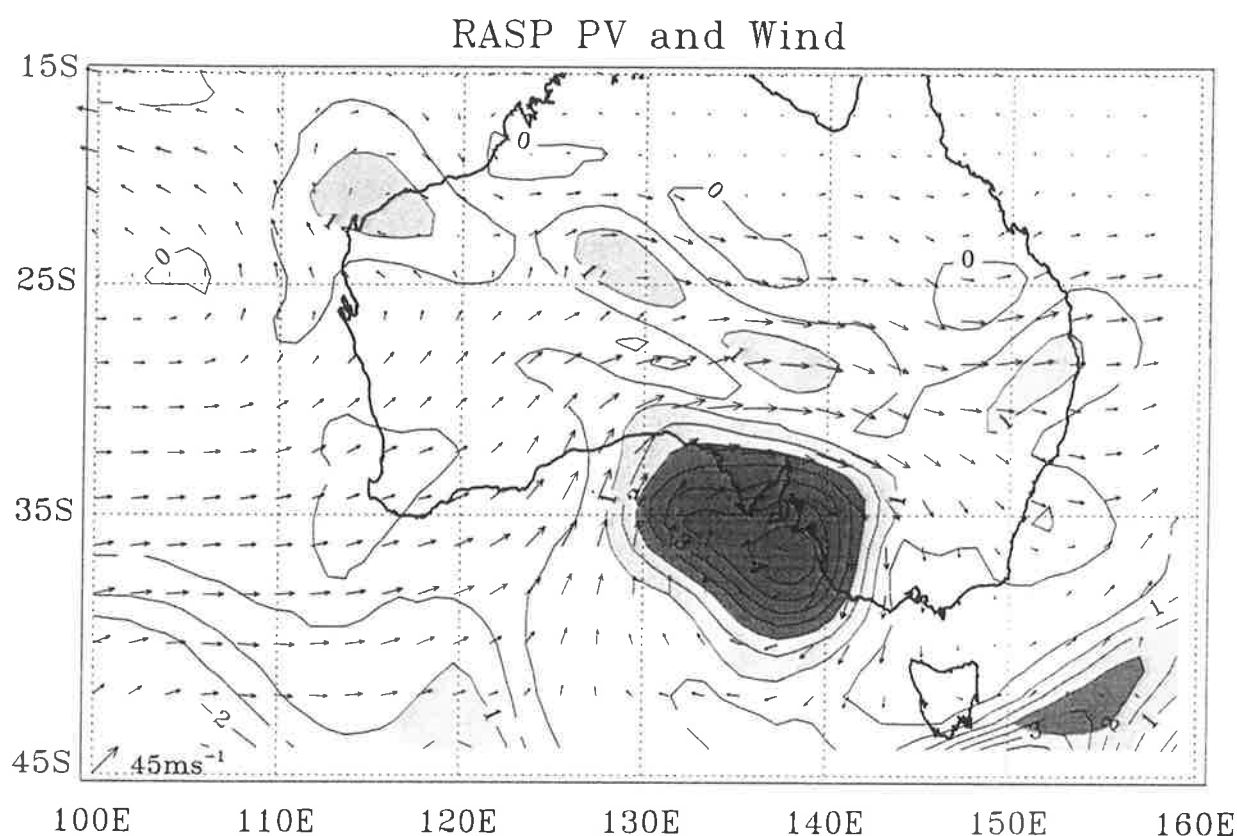


Figure 6.4: RASP potential vorticity (solid, contours every 0.5 PVU where 1 PVU is $10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$) and system-relative winds (vectorial, maximum wind reference vector in bottom left corner) on $\theta = 310 \text{ K}$ for 1100 UTC 29 August. Dark shaded areas indicate regions in which the surface is in the stratosphere, lighter shading indicates the tropopause.

overlying strong low-level temperature gradients (such as the situation displayed in Figure 6.2(a) and (b)) indicates development, exactly as we observe. However, the low-level θ gradients are weakening by 1100 UTC 29 August, and development would not be predicted: in accordance with this prediction, the surface low has weakened by 2300 UTC 29 August (Figure 6.2(d)). The evolution of the PV field is further discussed by *Mills & Wu* [1995].

Figure 6.5 shows the 300 hPa geopotential height and isotachs. A diffluent trough and strong upper-level jet streak are evident, with jet-core speeds exceeding 80 ms^{-1} . The surface low is located under the poleward side of the jet exit, and an upper-level ridge stretches along the east coast of Australia. *Mills & Wu* [1995] show that warm air advection at upper levels ahead of the tropopause fold is followed by cold air advection

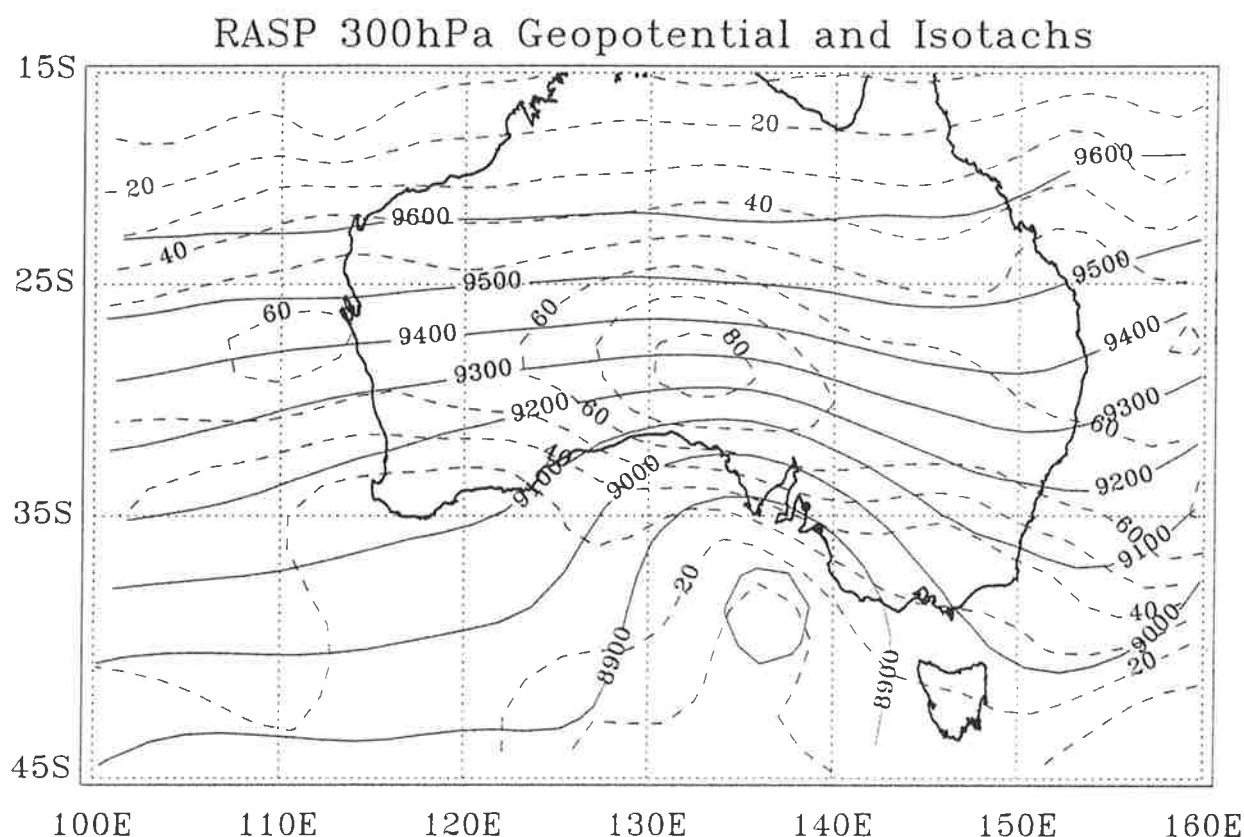


Figure 6.5: 300 hPa geopotential height (solid, 100 m intervals) and isotachs (dashed, 10 ms^{-1} intervals) from RASP for 1100 UTC 29 August.

behind, a pattern which often accompanies deep tropopause folding [Boyle & Bosart, 1986].

As already noted, the RASP analyses show strong low-level temperature gradients (the subtropical front) through central Australia associated with the North-west cloud band. The structure and evolution of subtropical fronts during late winter and early spring have recently been investigated by Smith *et al.* [1995], who found a consistent association with synoptic patterns and cut-off low developments similar to those presented here. Figure 6.6 shows a sequence of 850 hPa potential temperature maps, and is subtly different to Figure 6.2 where moisture was included to contour *equivalent* potential temperature. There are two distinct regions of temperature gradient: one through central Australia, which strengthens over the sequence, and the other associated with the low pressure system south of the continent, which weakens with time. These features retain separate identities throughout the sequence. Shaded contours

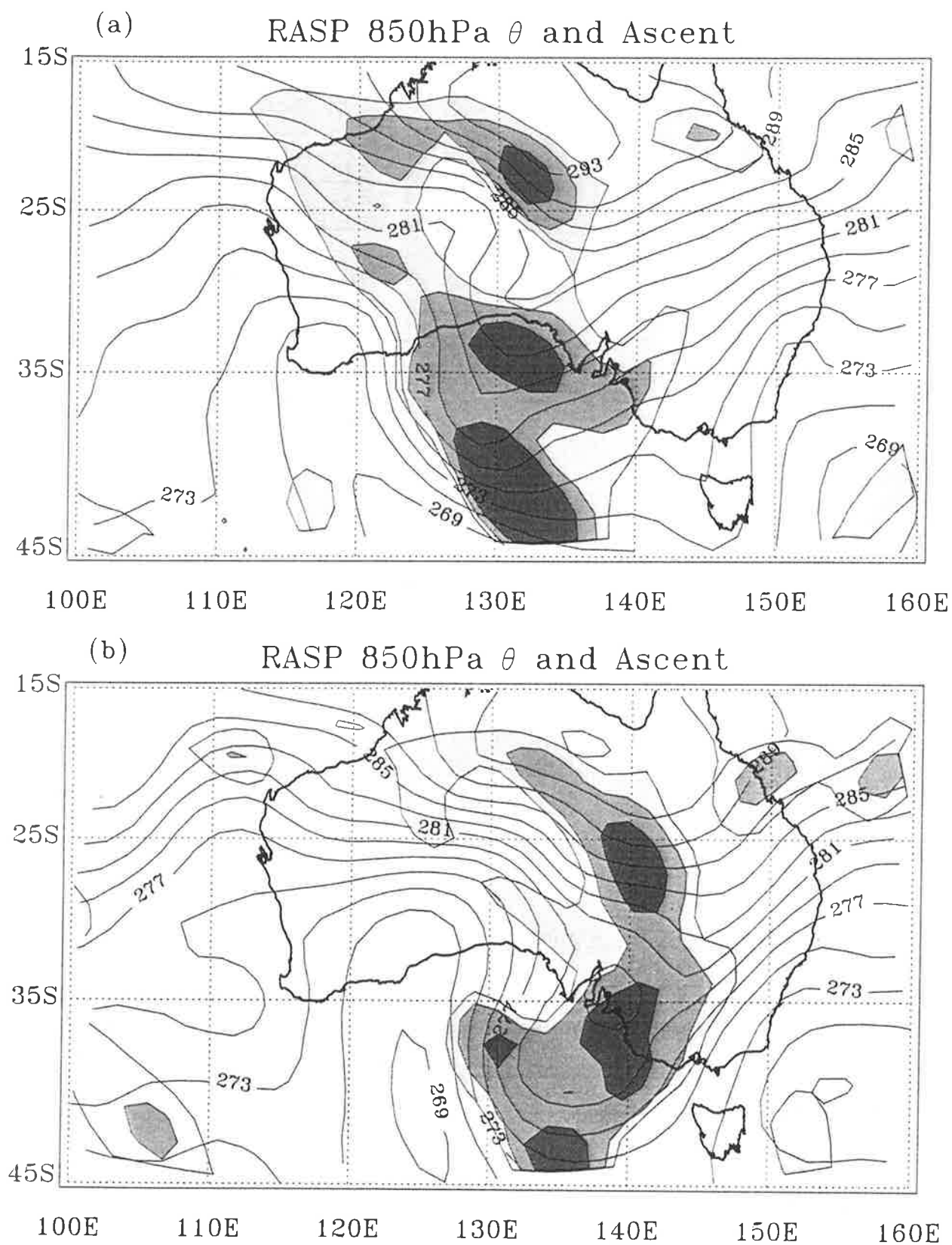
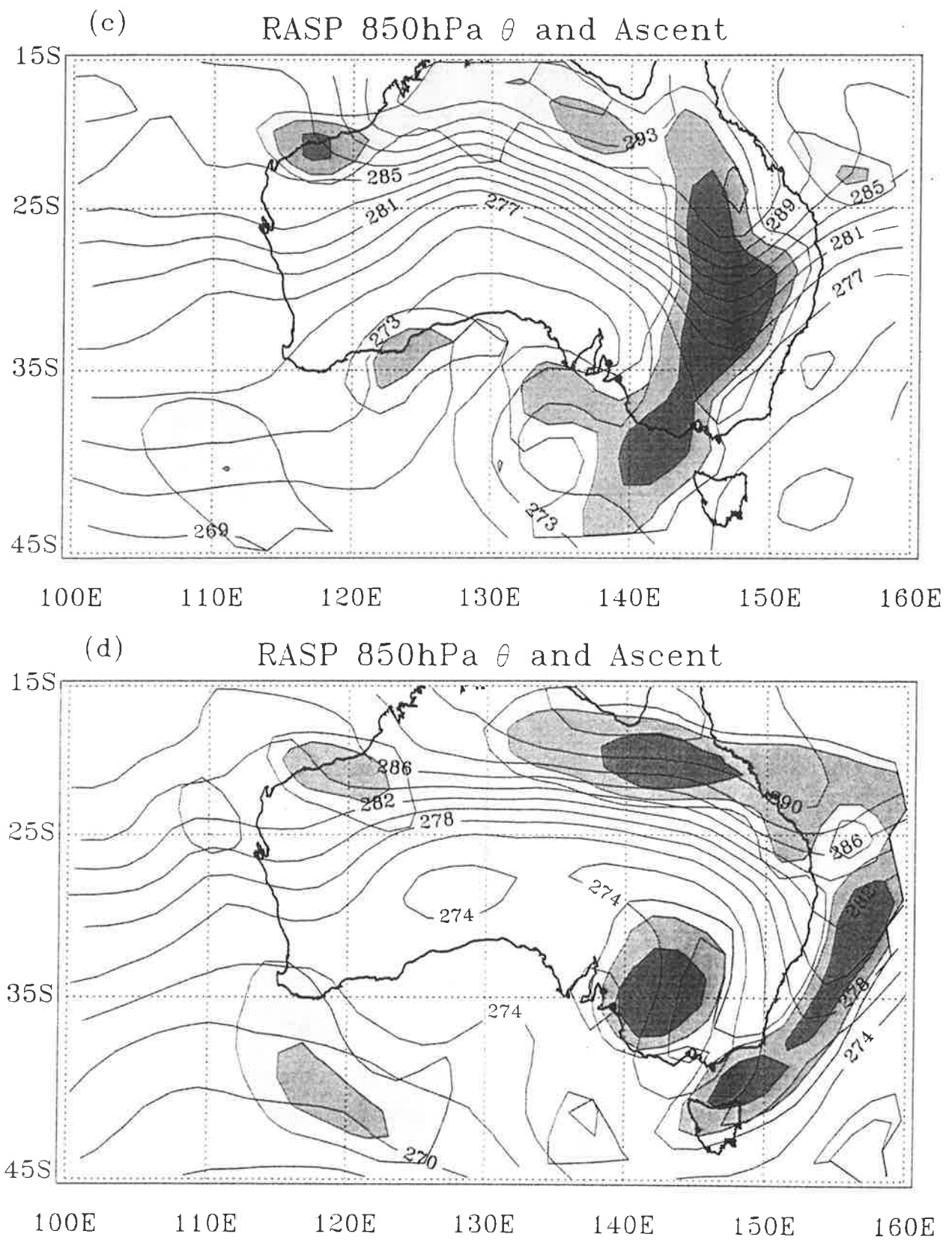


Figure 6.6: 850 hPa potential temperature (solid, 2K intervals) and ascent (shaded regions 10 [darkest], 5, and 2 [lightest] hPa hr⁻¹) from the RASP analyses at: (a) 1100 UTC 28 August; (b) 2300 UTC 28 August; (c) 1100 UTC 29 August; and (d) 2300 UTC 29 August.



of vertical motion (850 hPa omega velocities, or dp/dt , from RASP) are also plotted on each frame: the low-level ascent in the warm tongue and cloud head as they wrap around the surface low are readily apparent (compare with Figure 6.2, for example). *Griffiths* [1996] analyses frontogenesis in this region for this event via the generalised Petterssen frontogenesis function and F -vectors ($\mathbf{F} = \frac{D}{Dt} \nabla \theta$ where $\frac{D}{Dt} = \frac{\partial}{\partial t} + (\mathbf{u} \cdot \nabla)$) as defined by *Keyser et al.* [1988], and finds that the strongest frontogenesis is through the subtropics, consistent with *Smith et al.* [1995]. In addition, the region south of the continent is frontolytic, explaining the weakening of the temperature gradients in that region.

6.4 Mesoscale Observations

6.4.1 Surface Features

The most pronounced regional feature associated with the system was high rainfall, with many stations around Adelaide receiving their highest August daily rainfall on record. Lenswood, 20 km east of Adelaide received 70 mm of rain between 1000 UTC and 2300 UTC 29 August, including 17 mm in one half-hour period around 1930 UTC [*Furler*, 1993] when the minimum pressure was recorded in Adelaide (but after the low pressure centre to the south had started to move east). Surface wind gusts near 25 ms^{-1} were recorded around Adelaide during this time.

Automatic weather station data recorded at Buckland Park during the IOP are shown in Figure 6.7, with time running from right to left as an aid to comparisons with the cross-sections presented later. Pressure and temperature time series are displayed in Figure 6.7(a), while Figure 6.7(b) shows the zonal and meridional wind components (the anemometer was located at a height of 1.5 m). All time series have been smoothed by the application of a five-minute running mean. The minimum pressure occurs around 1900 UTC 29 August 1992, and is accompanied by wind gusts exceeding 12 ms^{-1} (Figure 6.7(c)) and the heaviest observed rainfall. At the same time, the temperature rises from a minimum of about 8°C at 1700 UTC 29 August (0230 LST 30 August) to nearly 14°C at 0300 UTC (1230 LST) 30 August, with the

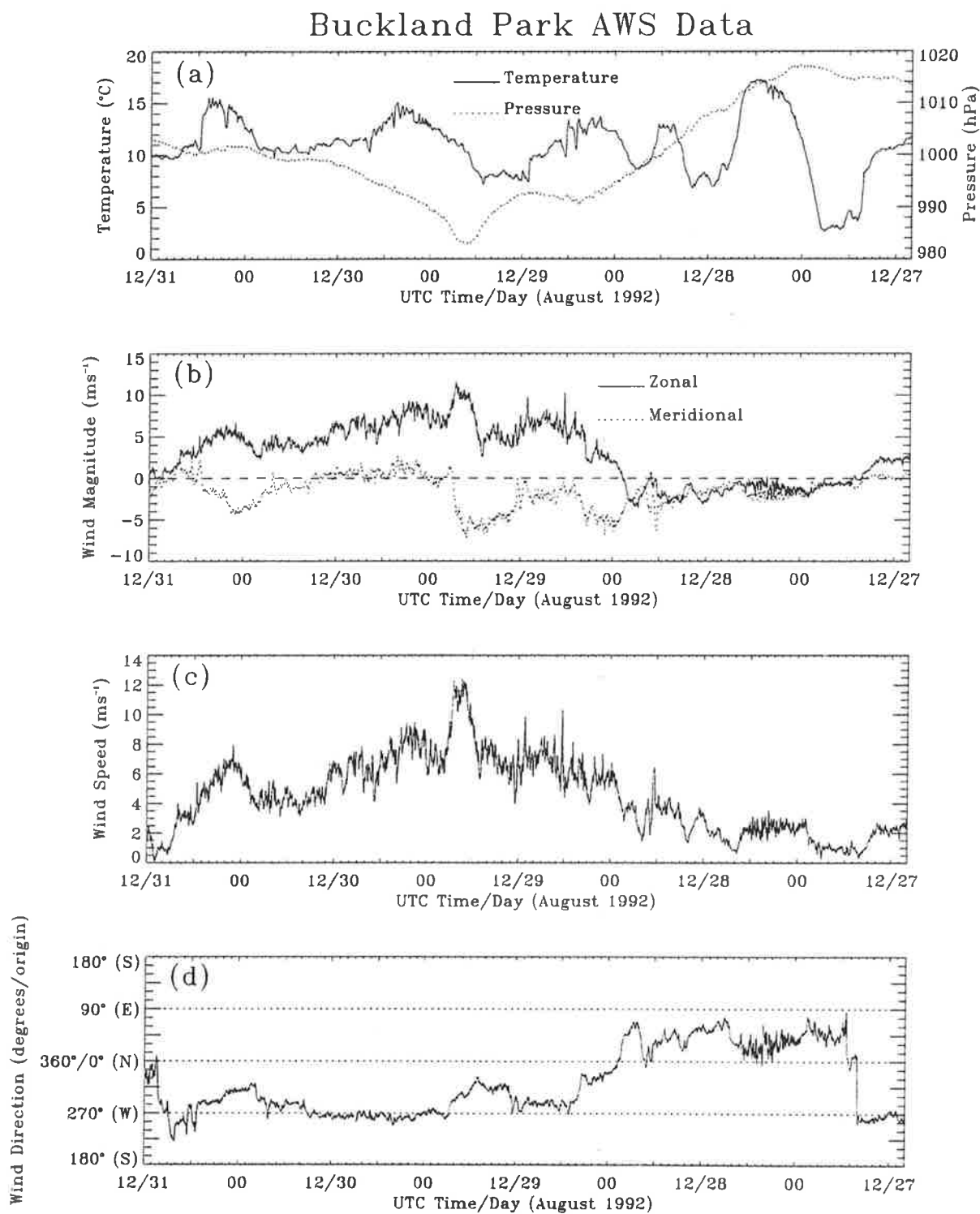


Figure 6.7: Buckland Park automatic weather station data during IOP IV/3: (a) temperature (solid, °C) and pressure (dotted, hPa); (b) zonal (solid) and meridional (dotted) wind components (ms⁻¹); (c) wind speed (ms⁻¹); and (d) direction of origin of wind.

most rapid increase in the early hours of the morning, against the diurnal trend; this is due to the advection of the warm tongue around the cut-off low (see Figure 6.2(c), for example). During the IOP, the surface wind changes direction from north-easterly, through northerly, to westerly (Figure 6.7(d)).

We note that a detailed surface analysis based on the South Australian Regional Office's surface station data and GMS infrared satellite imagery is consistent with RASP, with the expected wind change along the leading edge of the warm tongue [Griffiths, 1996]; that is, a "snapshot" of surface data at 50–100 km resolution around the South Australian coast is consistent with the ensemble of features we have outlined above.

6.4.2 Meridional Structure

Meridional cross-sections through 138°E at 1100 UTC 29 August from RASP analyses are shown in Figure 6.8. The dynamical tropopause, defined by the region between 1 and 2 PVU (where a PV unit is $10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$), is blackened in each panel. Note the change in tropopause height around 30°S, from about 15 km in the tropics to about 10 km in the midlatitudes. The definition of the dynamical tropopause causes this change to appear continuous, rather than the overlaying "break" feature common to the thermally-defined tropopause between the subtropics and midlatitudes. The perturbation ("fold") in the dynamical tropopause is consistent with that observed in other intense cyclones (see, for example, Hoskins & Berrisford [1988]).

Figure 6.8(a) shows contours of zonal wind. The winds are predominantly westerly throughout the period of observation, with a strong westerly jet evident near 11 km at 30°S; winds in the jet-core on the warm side of the fold exceed 70 ms^{-1} . The flow changes to easterly poleward of 35°S, reflecting the cyclonic circulation accompanying the tropopause fold. The secondary low-level wind maximum near 33°S and the strong low-level easterlies further to the south are part of the low-level cyclonic circulation around the cut-off low, which has its centre poleward of the fold. This wind structure shows great similarity to that shown in Browning & Hill [1985].

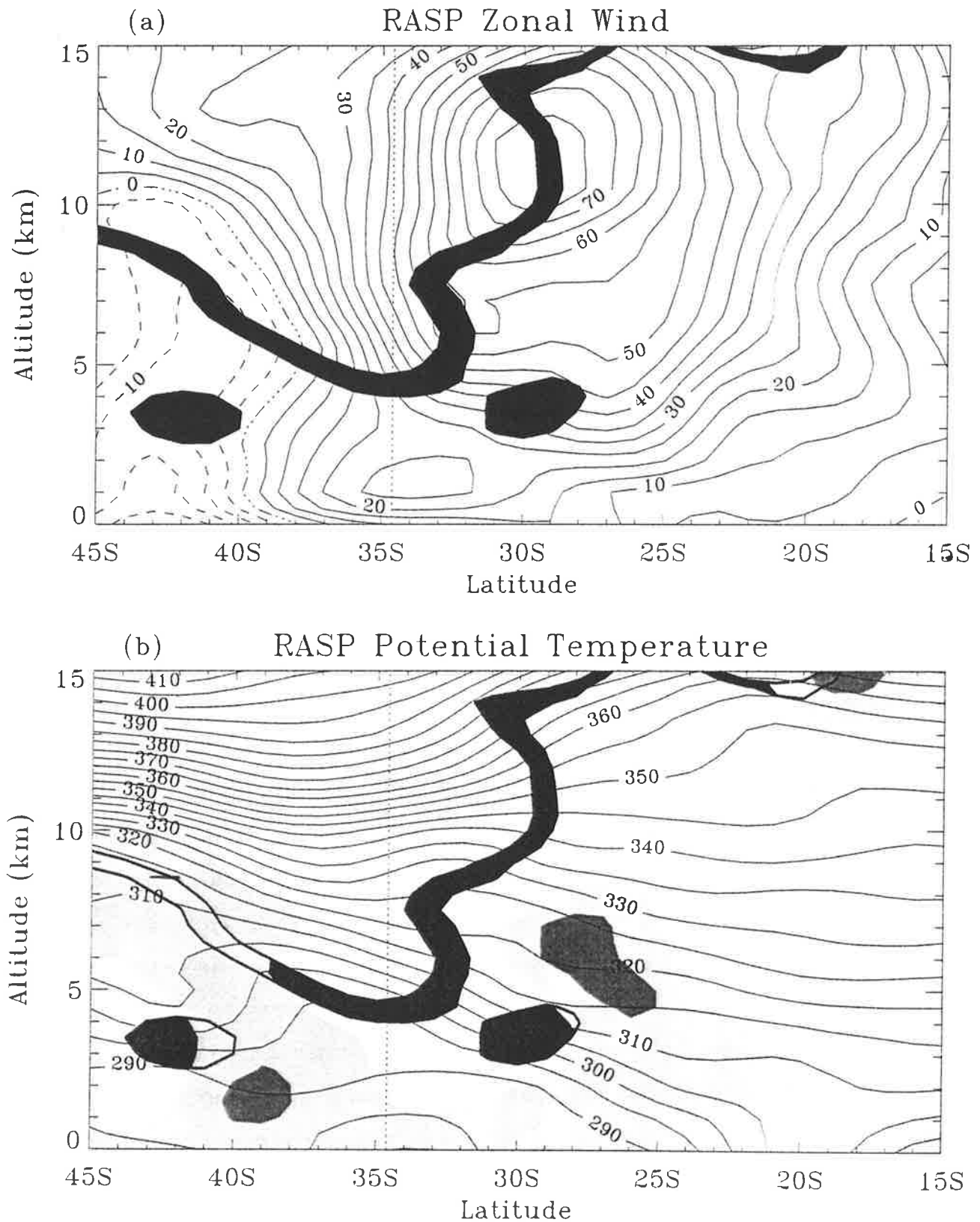


Figure 6.8: Meridional cross-sections through 138°E for IOP IV/3 at 1100 UTC 29 August from RASP: (a) zonal wind (5 ms^{-1} intervals); (b) potential temperature (solid, 5 K intervals) and relative humidity (shaded at 60% and 90%). The dynamical tropopause (between 1 and 2 PVU) is blackened, and the latitude of Adelaide is marked by a vertical dotted line.

Potential temperature is contoured in Figure 6.8(b), together with relative humidity. The concentration of isentropes at the surface near 23°S defines the subtropical front, sloping poleward to around 33°S at the tropopause fold. There is a cold anomaly at the surface under the fold near 35°S , while two deep moist layers are prominent features. The moist layer sloping polewards along the subtropical front is associated with the North-west cloud band, while the layer to the rear of the fold is along the warm tongue, marking the cut-off low cloud on the poleward side of the low (cloud band **C** in Figure 6.3). The dry slot between these two moist layers above the cold anomaly in the region of the low-level wind maximum seen in Figure 6.8(a) corresponds to the cold tongue discussed earlier.

6.4.3 Zonal Structure

Since the motion of weather systems through Adelaide is generally from west to east, it is possible to make a reasonably direct comparison between zonal cross-sections from RASP, and time-height cross-sections from the radiosonde and profiler data. However, some caution must be exercised in such a comparison, as we have discussed earlier. For example, the weather systems evolve over time, so features further away from the observing site (in both time and space) would be expected to display some differences. Additionally, there are considerations with regards to sonde-profiler comparisons, as we have outlined in Chapters 3 and 4. The radiosonde is a Lagrangian sensor, sampling along a trajectory which varies with the local wind and usually terminates 100km or more away from the point of release; each data point along the trajectory represents a spot measurement at that point in space and time. Conversely, the profiler measures a volume average (the shape of which can be approximated in our case by a cylinder of diameter 300m and height 500m) at each range gate along a purely vertical profile (an Eulerian measurement). The loss in vertical resolution with the profiler is compensated by a better estimate of the background, with fewer small-scale, transient effects. Averaging the profiler data in time gives a good picture of the evolution of the atmosphere, and allows for better time-space comparisons.

The system in question moves zonally at about 10 ms^{-1} (slowing to about 6 ms^{-1}

as it passes through Adelaide), or about 10° in longitude per day. The radiosonde and profiler cross-sections presented later cover about two days either side of the key time (1100 UTC 29 August, when the fold is over Adelaide), so the RASP cross-sections between about 120°E and 160°E are the most sensible regions for the purposes of comparison. The RASP zonal cross-sections are taken when the tropopause fold is over Adelaide, and we would expect the least distortion (due to the evolution of the system) in comparing different observing methods around that time (circa 1100 UTC 29 August). The time-height cross-sections derived from the radiosonde and profiler data are plotted with time running from right to left, so that a direct visual comparison may be made with the space-height RASP cross-sections. Also, note that the altitude ranges displayed are slightly different for each data source; although this makes overlay-type comparisons more difficult, we feel that it is better to display data on the most appropriate scale, thereby avoiding excess whitespace or the loss of useful data.

Figure 6.9 shows zonal cross-sections of θ and θ_e from RASP. Figure 6.10 shows time-height cross-sections of the same fields, but using thermodynamic data from the SAFEX sonde flights. The overall comparison is excellent, with all major features reproduced. The figures show a deep tropopause fold, with the tropopause raised in the warm air ahead of the surface low and lowered in the colder air behind. The analysed fold extends down to 5 km, with an intrusion of stratospheric (high $d\theta/dz$) air into the troposphere. At this time, we also note the region of air, apparently of tropospheric origin (uncharacteristically low $d\theta/dz$), in the fold above the stratospheric intrusion. We shall later show that this is the result of circulation through the fold. Note how the isentropes in the troposphere curve towards the upper anomaly, which those in the stratosphere bend downwards, in the manner described by *Hoskins et al.* [1985] and *Thorpe* [1986]. Numerical studies by *Griffiths & Reeder* [1995] and *Reeder & Griffiths* [1995], in which a jet front system is forced with both horizontal deformation and vertical shear, predict a similar tropopause fold to that seen here.

The low-level structure is that of a cold-core cyclone and it is typical of a cut-off low. Well ahead of the surface low, the air is dry and stable. The deep moist layer associated with the warm tongue east of the surface low, with its sharp leading edge near 1400 UTC 28 August consistent with the sharp southern boundary of the

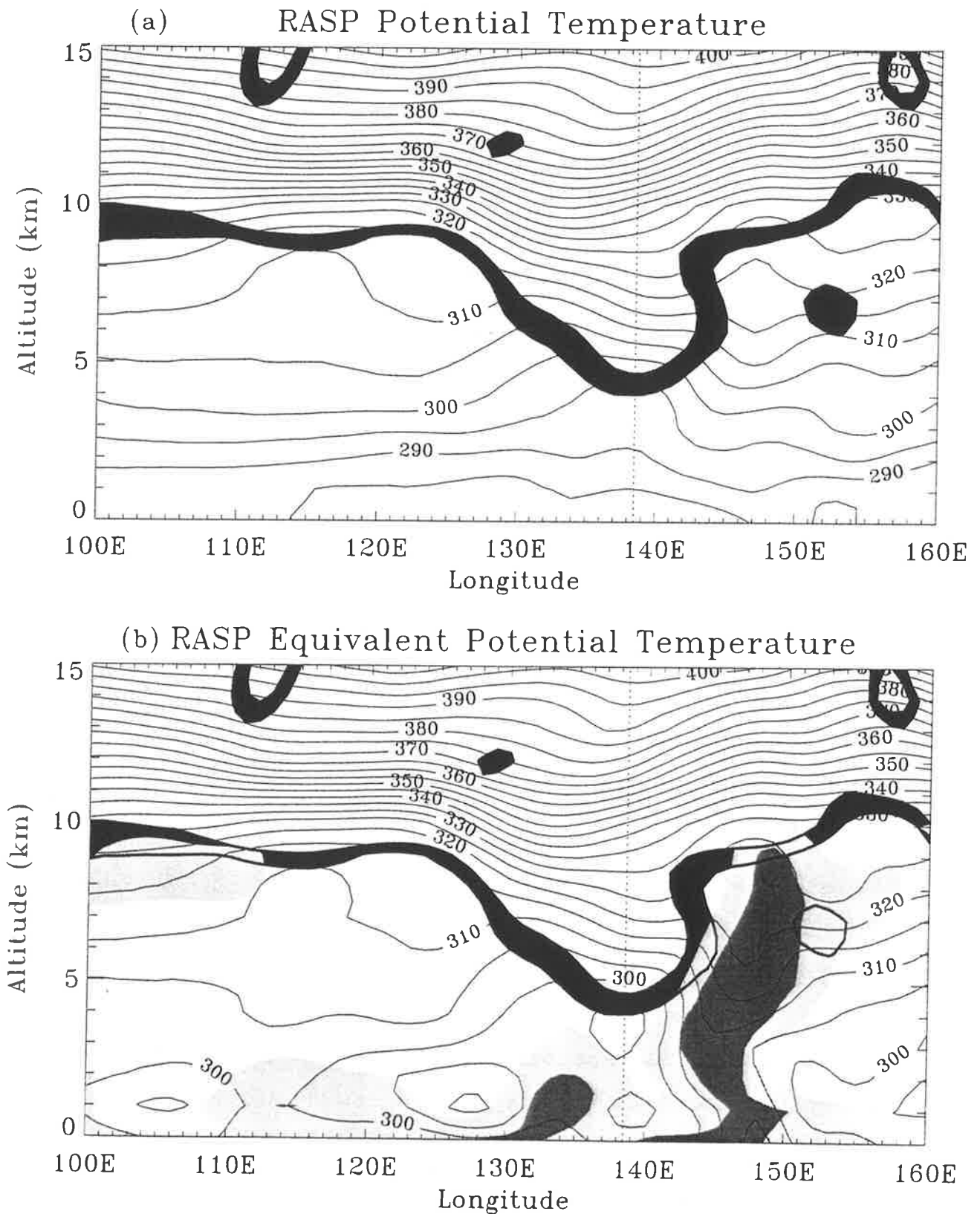


Figure 6.9: Zonal cross-sections through $34^{\circ}21'S$ for IOP IV/3 at 1100 UTC 29 August from RASP: (a) potential temperature (5 K intervals); (b) equivalent potential temperature (solid lines, 5 K intervals) and relative humidity (shaded at 60% and 90%). Tropopause and Adelaide marked as in Figure 6.8.

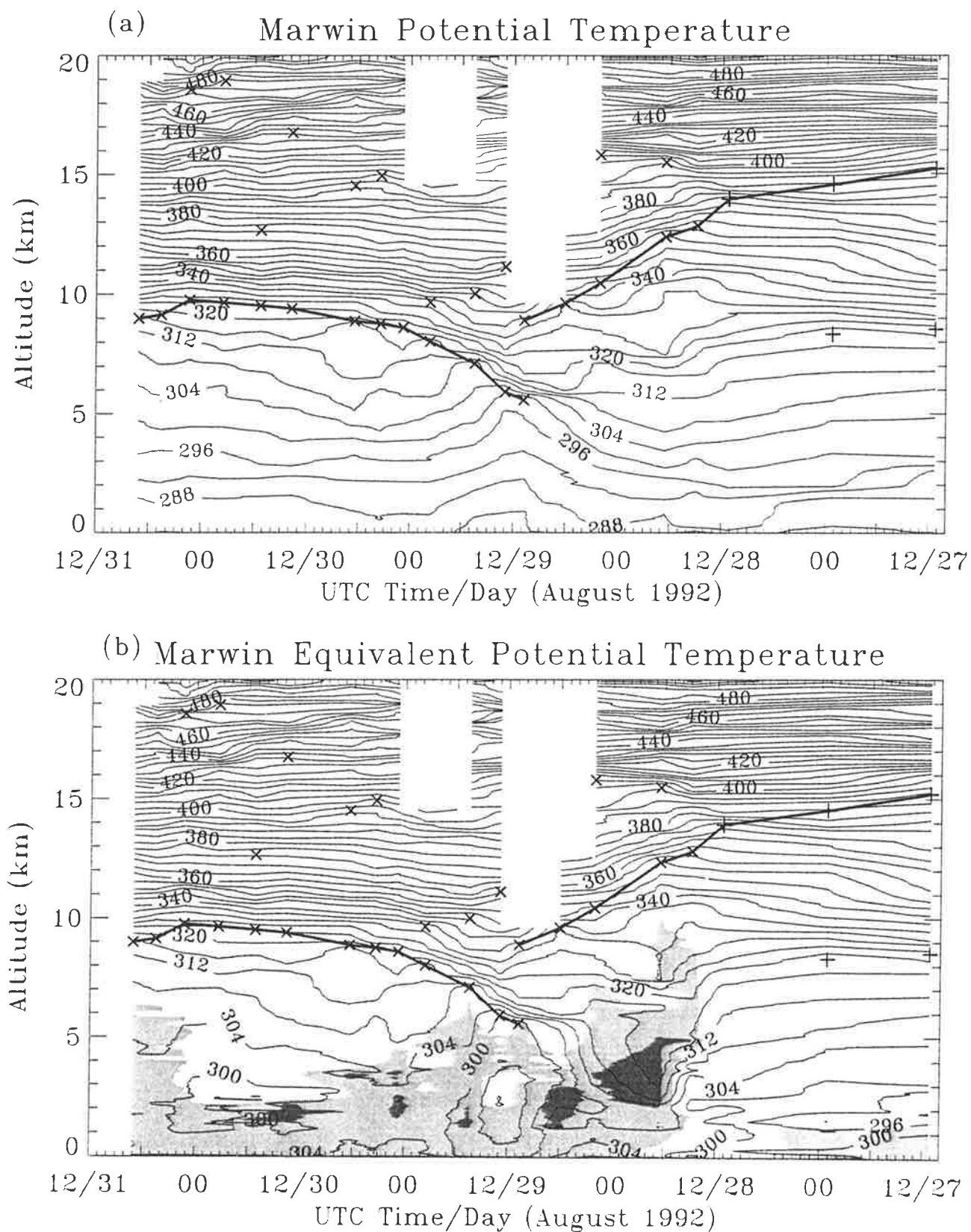


Figure 6.10: Time-height cross-sections for IOP IV/3 from SAFEx radiosonde data: (a) potential temperature (4K intervals); (b) equivalent potential temperature (solid lines, 4K intervals) and relative humidity (shaded at 60% and 90%). The sonde-determined tropopause heights are marked by $\times$'s (Adelaide airport flights by $+$), joined by a heavy solid line.

North-west cloud band in Figure 6.3, is marked by a warm-frontal-like gradient in θ_e and humidity, while the trailing edge (near 0300 UTC 29 August) coincides with the upper cold front and tropopause fold. The dry slot and cold tongue are seen in both RASP and radiosonde analyses. The air below about 5 km is well mixed after about 0600 UTC 29 August, with regions of potential instability extending well into the cold air. The trailing edge of the midlatitude tropopause can be seen in the first twelve hours of the MARWIN analyses, indicating the “unperturbed” state of the atmosphere over Adelaide, with tropopause-like gradients near 15 km and 10 km altitudes. Note also the adiabatic regions around 9 km at 1800 UTC 28 August and 8 km at 0600 UTC 30 August where (as we have seen earlier in the thesis, and will see later in this chapter) the profiler experiences “biteout”.

Figures 6.11, 6.12 and 6.13 present zonal (or time-height) cross sections of the horizontal wind field from RASP, MARWINsonde, and profiler (spaced antenna, two-hour average) data respectively. Note the different altitude ranges in each figure, reflecting the capabilities of each data source. The winds are predominantly westerly (eastward) throughout the period of observation, with an upper-level zonal jet of over 80 ms^{-1} near 11 km on the warm side of the tropopause fold around 1500 UTC 28 August (Figure 6.12(a); or 150°E in the RASP cross-section, Figure 6.11(a)); this (subtropical front) jet decreases in strength with time. The sampling time problem outlined in Section 3.4 affects the profiler measurements around the time of the jet, although it is gratifying to note the tendency towards “missing data” rather than erroneous returns. The profiler winds through the remainder of the IOP show good agreement with both RASP and sonde retrievals. As predicted, the profiler experiences “biteouts” in the returns through the regions of near-adiabatic lapse rate mentioned above; this is especially clear around 0600 UTC 30 August, where the missing data occurs in a region of otherwise high data density. It is instructive to note that the RASP data are generally of lower magnitude, due to averaging over the 150 km, 6 hour grid, while the MARWIN data are generally of higher magnitude due to the tendency of the sonde to accentuate “gusts” throughout a flight (although some averaging in the MARWIN software moderates the latter effect).

The cyclonic structure of the system is obvious in the meridional wind fields (panel

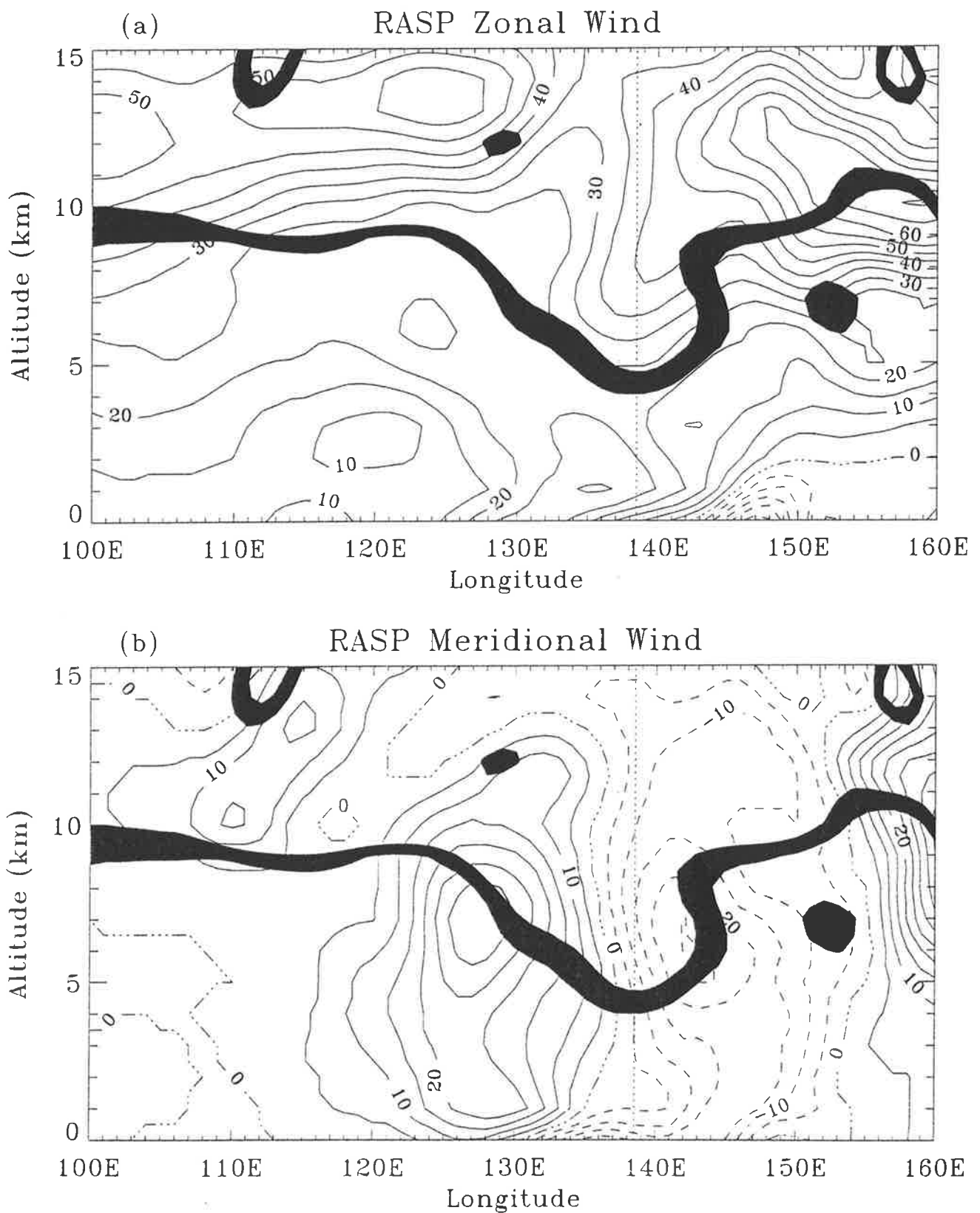


Figure 6.11: Zonal cross-sections through $34^{\circ}21'S$ for IOP IV/3 at 1100 UTC 29 August from RASP: (a) zonal wind; and (b) meridional wind (5 ms^{-1} intervals, positive values solid, zero-contour triple-chained, negative values dashed). Tropopause and Adelaide marked as in Figure 6.8.

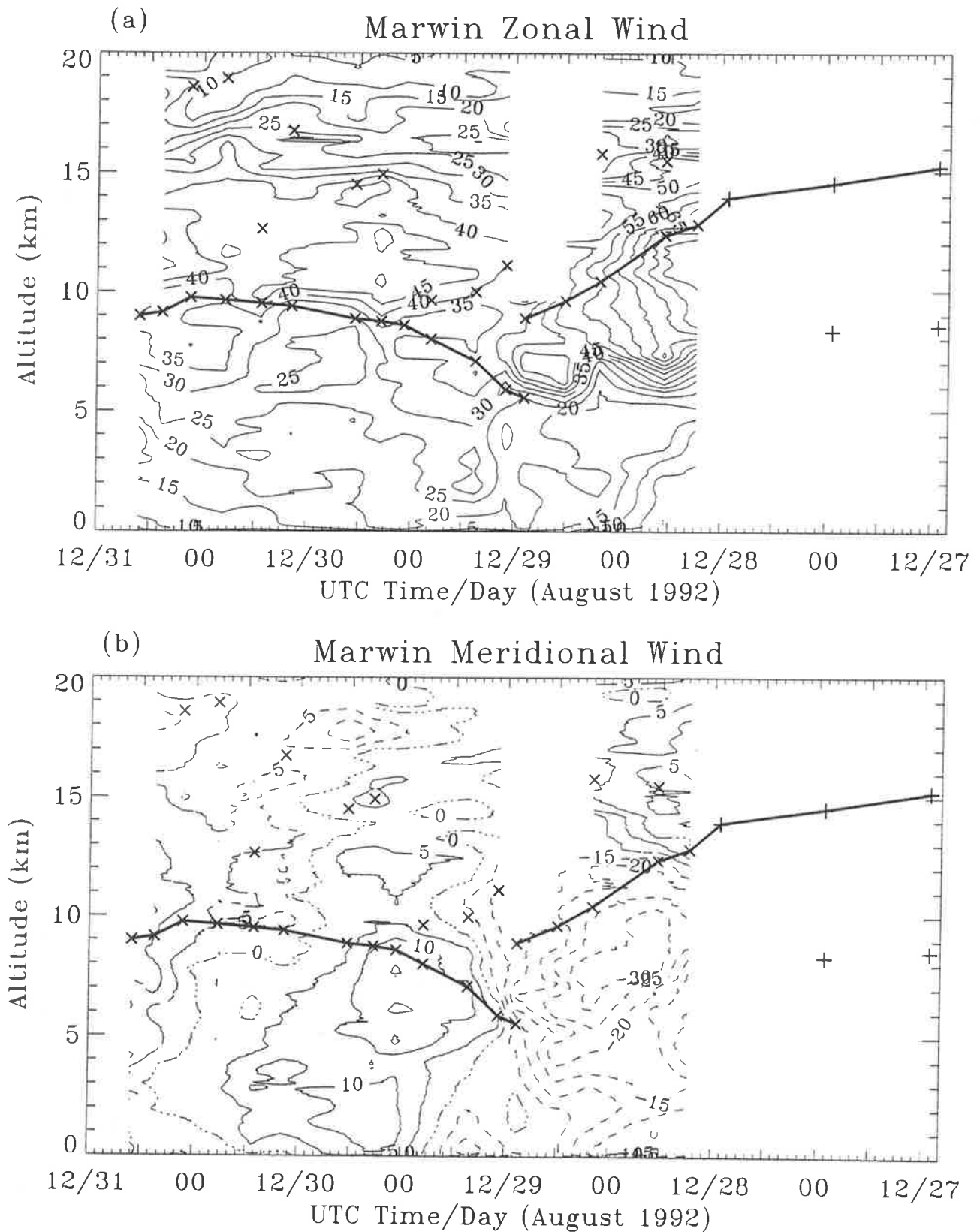


Figure 6.12: Time-height cross-sections for IOP IV/3 from SAFEx radiosonde data: (a) zonal wind; and (b) meridional wind (5 ms^{-1} intervals, positive values solid, zero-contour triple-chained, negative values dashed). Tropopause markers as for Figure 6.10.

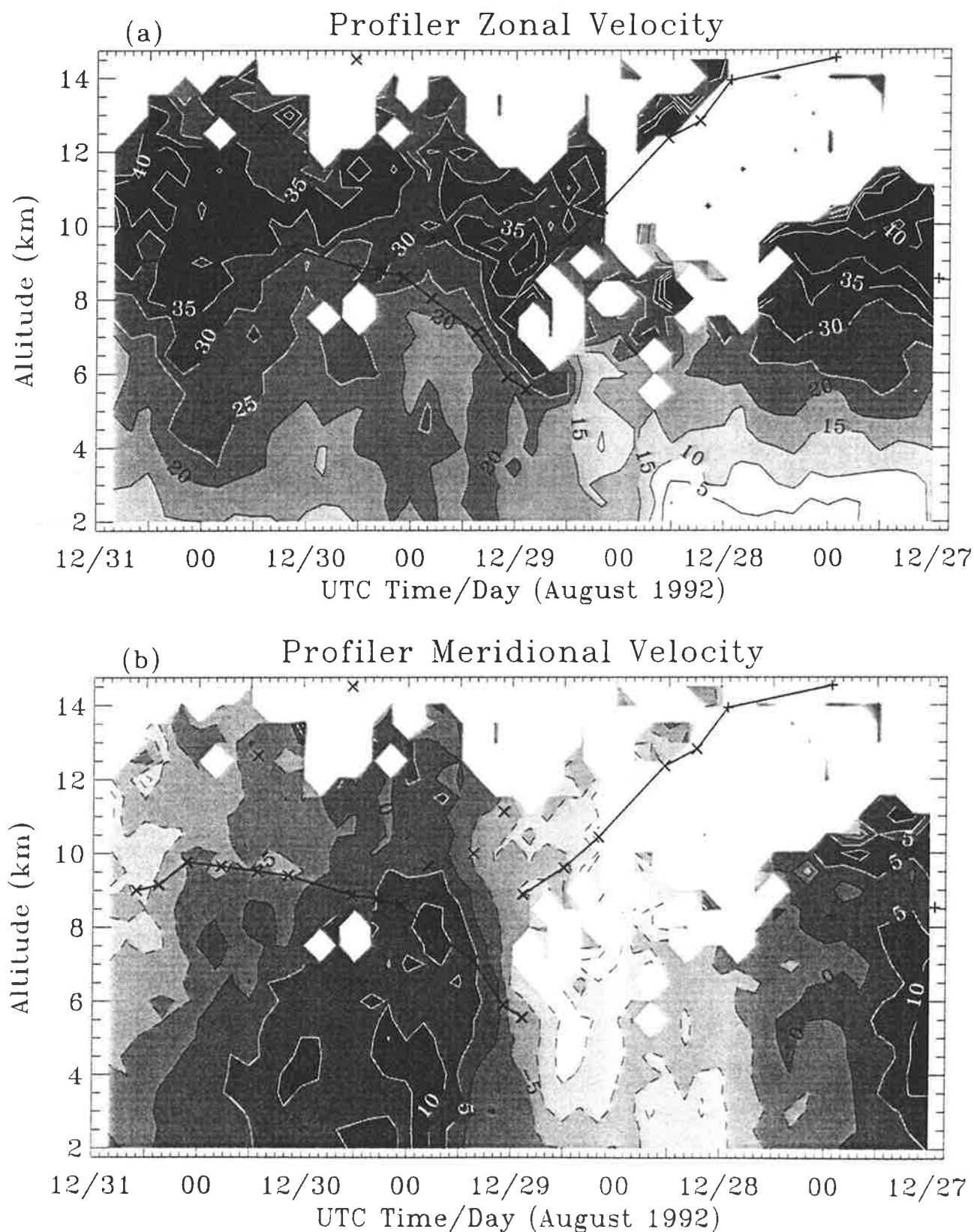


Figure 6.13: Time-height cross-sections for IOP IV/3 from profiler spaced-antenna data: (a) zonal wind; and (b) meridional wind (two-hour average: 5 ms^{-1} intervals, positive values solid, zero-contour triple-chained, negative values dashed). Tropopause markers as for Figure 6.10.

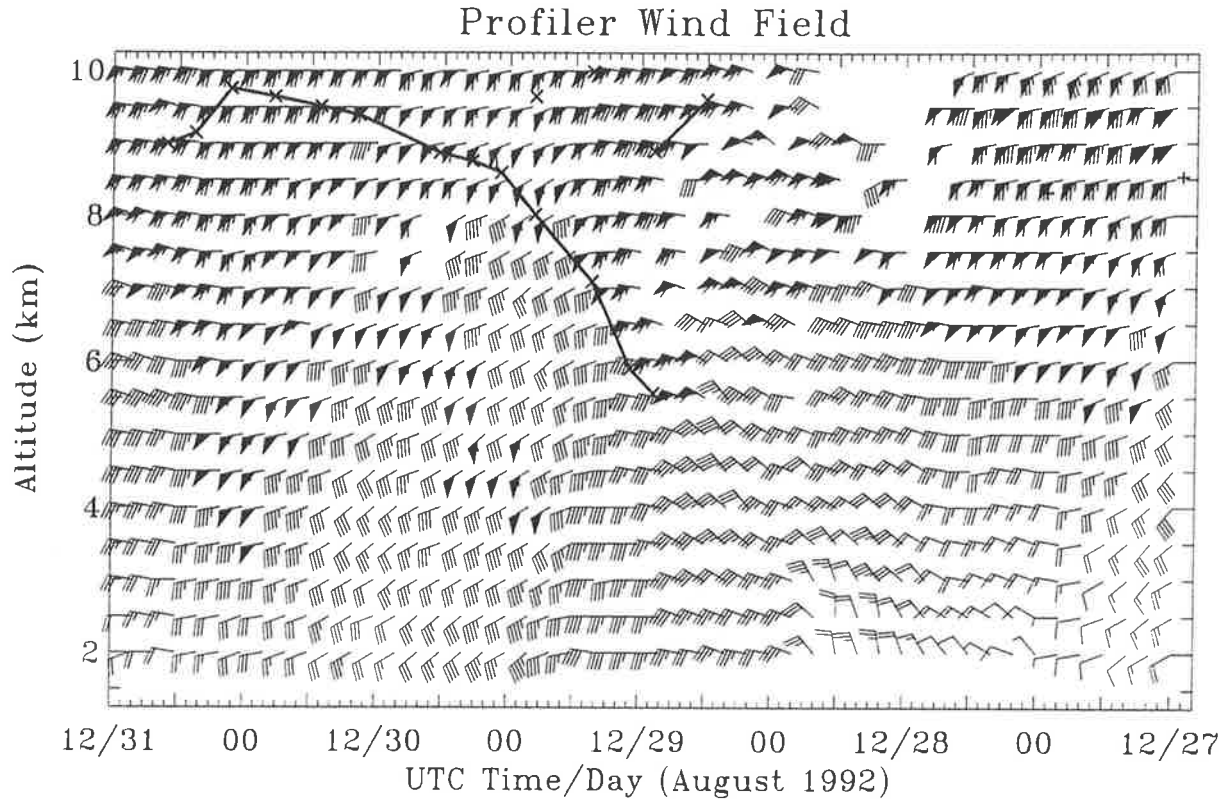


Figure 6.14: Time-height cross-section of the profiler spaced-antenna wind field, displayed as windbarbs (flag 25 ms^{-1} , barb 5 ms^{-1} and half-barb 2.5 ms^{-1} , at the “tail” of the wind-vector). Tropopause markers as for Figure 6.10.

(b) in each of the preceding three figures), with northerly (southward) flow ahead of the system and southerly (northward) flow behind. A region of particularly strong cyclonic shear ($\partial v / \partial x$) at 6 km altitude occurs at about 1400 UTC 29 August in conjunction with the tropopause fold. The profiler wind components are combined in Figure 6.14 over a reduced vertical range. The circulation on either side of the fold is clear, although the wind change is not as marked as would be expected for a classical cold-frontal system.

Figure 6.15 shows isotachs of two-hour average vertical velocity as measured by the profiler. Panel (a) displays the raw radial velocities derived from the vertical Doppler beam. However, as we have discussed in Section 3.1.2, if the beam is not perfectly vertical, then the radial velocities will be contaminated by part of the horizontal wind field “folding” into the radial estimate. The correction to the radial velocities is geometrically simple, provided one knows the actual beam direction. For example,

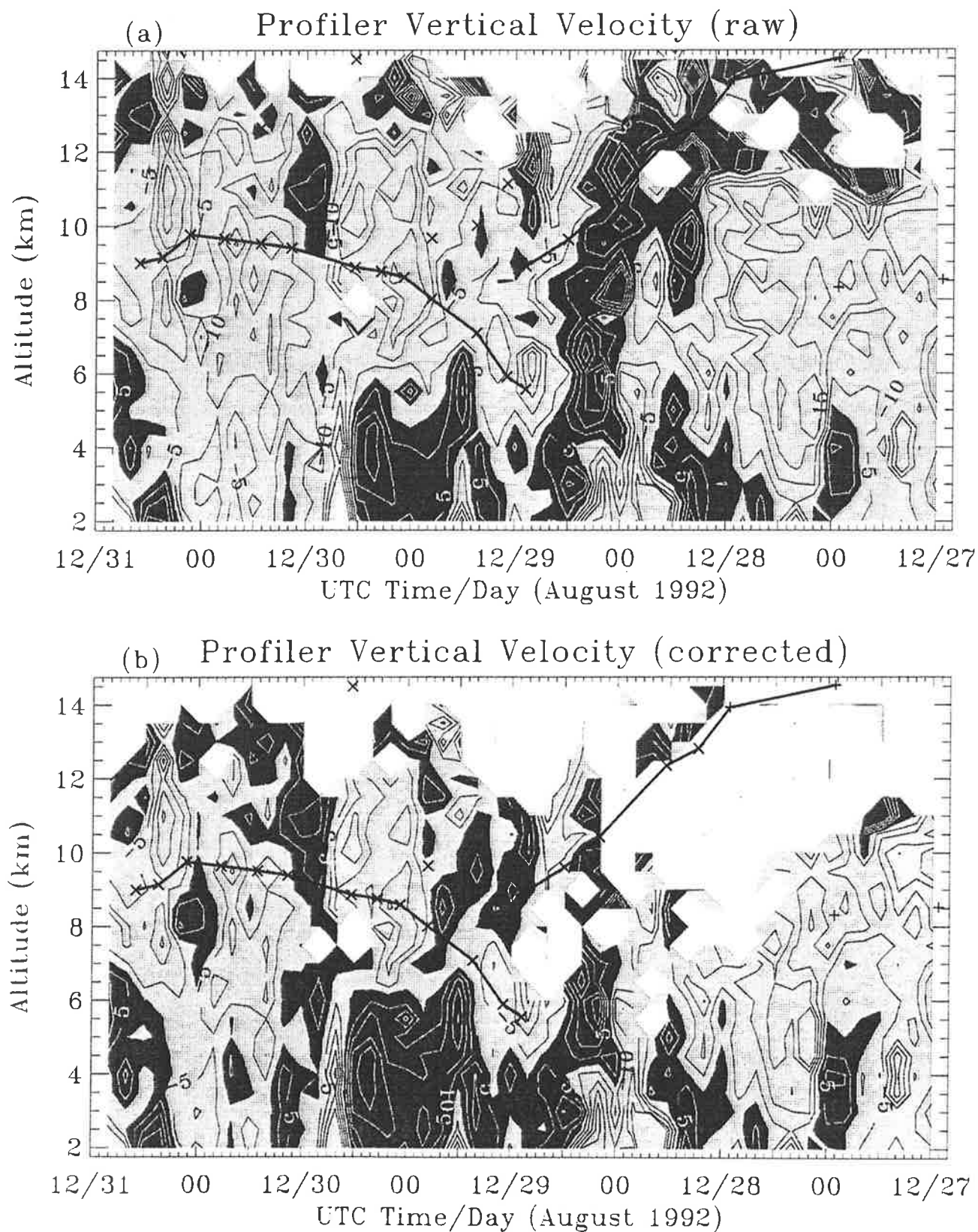


Figure 6.15: Time-height cross-section of profiler two-hour average Doppler vertical velocity (5 cm s^{-1} contour intervals, regions of ascent shaded darker): (a) raw “vertical” beam retrievals; and (b) retrievals corrected for contamination by the horizontal wind due to beam tilt and phasing (see Section 3.1.2). Tropopause markers as for Figure 6.10.

if the beam points towards (χ, ψ) in spherical coordinates (χ is the angle from zenith and ψ is the pointing azimuth), then the “true” vertical velocity w can be derived from the radial estimate w_r in a regime where the horizontal velocity $\mathbf{u}_h = (u, v)$ by:

$$w = \frac{w_r}{\cos \chi} - u \sin \psi \tan \chi - v \cos \psi \tan \chi \quad (6.1)$$

In our case, measurements of the physical array tilt give $(\chi, \psi) = (0.18^\circ, 210^\circ)$ (A. Kazempour, private communication 1996), which can be combined with our estimates of the phasing errors to produce the “corrected” vertical velocities displayed in Figure 6.15(b). In simple terms, the correction is approximately 5 cm s^{-1} (additive), or about one contour level. Note how the correction has balanced (in broad terms) regions of ascent and subsidence, rather than the unrealistic widespread subsidence (due to positive zonal velocities causing a negative contribution from folding on the slightly westward-pointed beam) seen in Figure 6.15(a).

The regions of strong descent near 1800 UTC 29 August at the lowest range gates is due to hydrometeor contamination (rain echoes); the remainder of the diagram indicates air motions. Although the possibility of some small negative bias due to unrecorded rainfall should be borne in mind, there was certainly no other long or intense rain period at the field site (see the discussion regarding Figure 6.17 later). There is a deep region of ascent in the upper warm air ahead of the fold (0000 UTC to 0600 UTC 29 August) and in a broader region in the cold air behind (1800 UTC 29 August to 0600 UTC 30 August). Air subsides prior to about 0000 UTC 29 August, through the tropopause fold, and throughout most of the cold air from 0600 UTC 30 August onwards. If we compare this pattern with the moisture distribution (shaded on Figures 6.9 and 6.10, for example), we see ascent in moist regions, and subsidence in dryer regions. This is consistent with the dry slot being cold dry air subsiding, while moist air ascends via the conveyor belts outlined earlier.

RASP-derived vertical velocities at our standard 1100 UTC 29 August comparison time (when the fold is over Adelaide) are displayed in Figure 6.16. We note an excellent comparison in terms of the location of regions of ascent and subsidence (recalling the difference between horizontal scales mentioned earlier with regard to comparisons between time and space cross-sections), although the profiler magnitudes are larger.

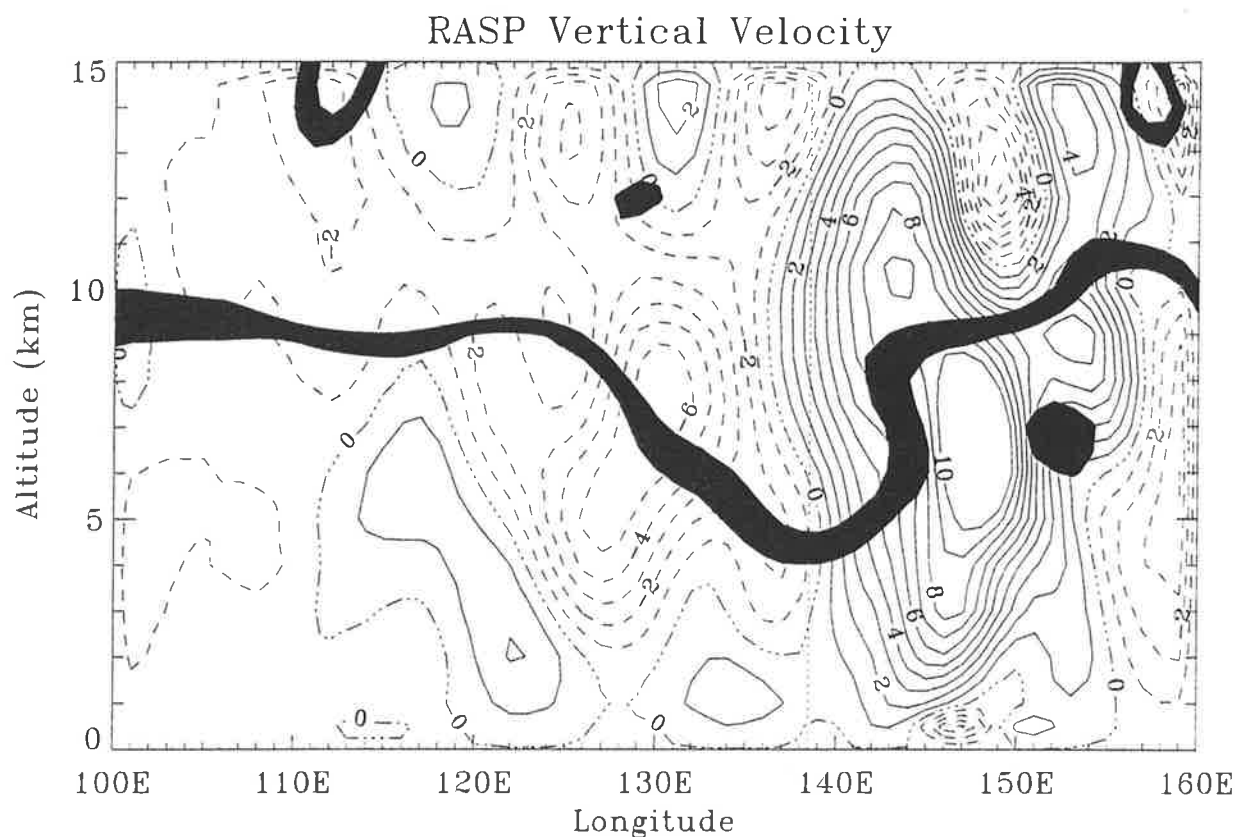


Figure 6.16: Zonal cross-section of vertical wind through $34^{\circ}21'S$ for IOP IV/3 at 1100 UTC 29 August from RASP. Contours are at 1 cm s^{-1} intervals: positive contours indicating ascent are solid, negative contours (subsidence) are dashed, and the zero-contour is triple-chained. Tropopause and Adelaide marked as in Figure 6.8.

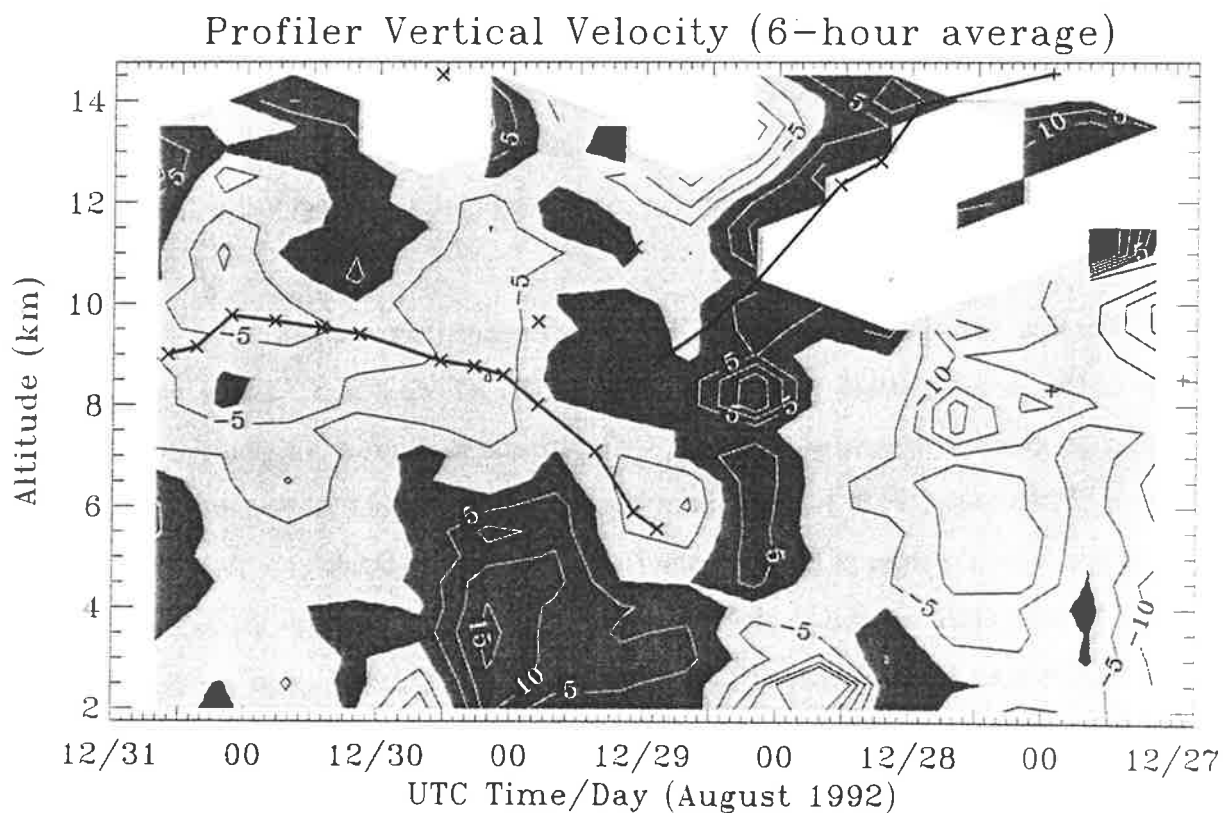


Figure 6.17: Time-height cross-section of profiler six-hour average Doppler (corrected) vertical velocity (5 cm s^{-1} contour intervals, regions of ascent shaded darker), for comparison with RASP analyses. Tropopause markers as in Figure 6.10.

This is expected, since the RASP analyses involve averages over such a large region in space (150 km grid) and time (six-hourly model realisations). In order to make a more representative comparison, we show the profiler vertical velocities matched (via time averaging) to the RASP time resolution of six hours in Figure 6.17. The magnitudes are in much closer agreement with the RASP analyses, especially in the pre-fold and fold regions. There is a magnitude discrepancy in the cold air immediately behind the fold (after 1800 UTC 29 August), but this is not unexpected given the generally rapid evolution of the vertical velocity field, and especially as we move away from the time and place where the RASP analysis is most valid. Importantly, the pattern of ascent and subsidence is consistent between these independent techniques. There are very few independent verifications of profiler vertical velocity measurements (see *Larsen et al.* [1988] and *Nastrom et al.* [1994] as examples of previous work), and a good comparison in an evolving mesoscale system such as the present study is a significant contribution to promoting faith in profiler measurements. As a final point, note the pattern of subsidence along the lower edge of the fold, and ascent in the warm air ahead and on the upper edge. This pattern is readily apparent in the six-hour averages and implies a circulation, or stratosphere-troposphere exchange of air, through the fold: stratospheric air entering the troposphere along the lower boundary, while air of tropospheric character mixes into the stratosphere aloft. Recall the discussion of Figure 6.10, where the potential temperature contours in this region showed stratospheric and tropospheric character in separate regions of the fold.

Another useful indicator of the difference between tropospheric and stratospheric air is the Brunt-Väisälä frequency, $N^2 = (g/T)(dT/dz + \Gamma_d)$. Large values of N^2 are characteristic of stratospheric air, with altitude profiles of N^2 generally showing a jump from about $1 \times 10^{-4} \text{ s}^{-2}$ to about $4 \times 10^{-4} \text{ s}^{-2}$ across the tropopause. Figure 6.18 displays derived values of N^2 from the Buckland Park and Adelaide Airport launches, with larger values shaded darker. The intrusion of stratospheric air into the troposphere down to 5.5 km between 0300 UTC and 2100 UTC 29 August is clear, as is the raised tropopause in the warm air and the near-adiabatic regions in the upper troposphere (1800 UTC 28 August and 0600 UTC 29 August) where the profiler experienced

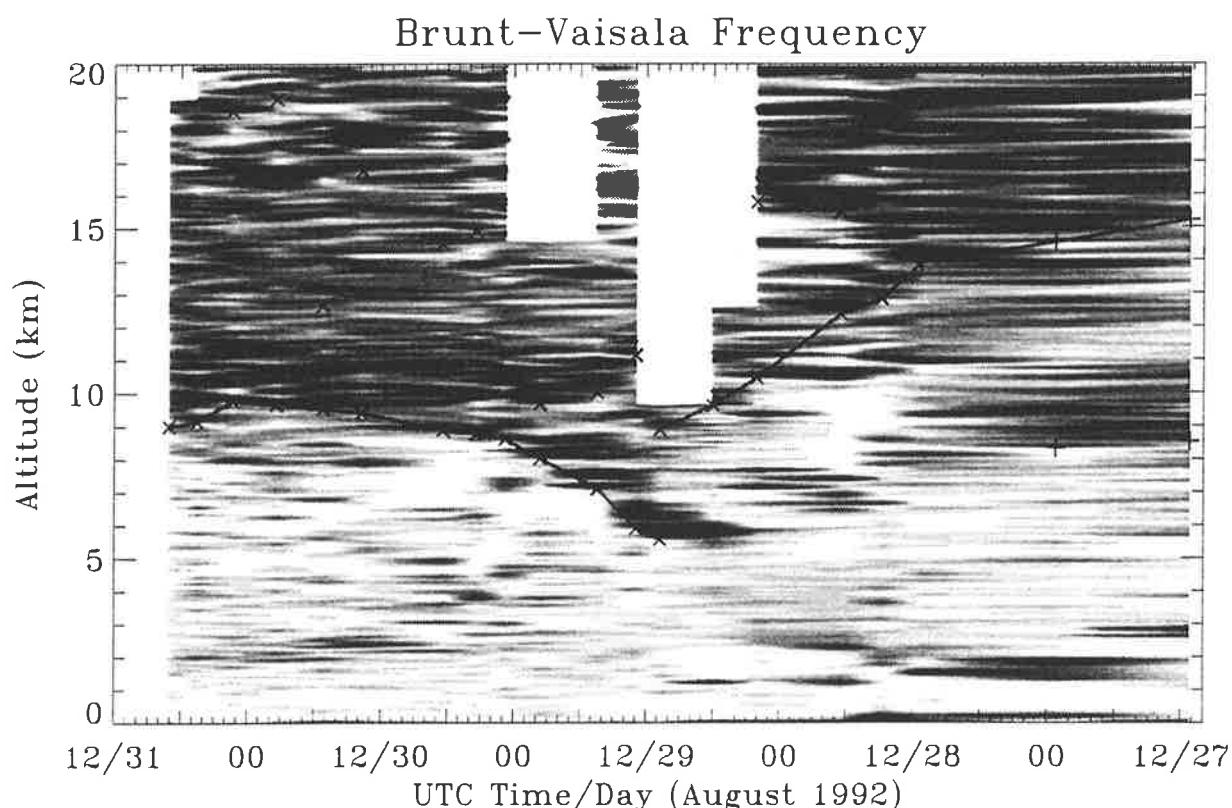


Figure 6.18: Time-height cross-section of N^2 for IOP IV/3, derived from Buckland Park sonde data. Larger values of N^2 are shaded dark; the dynamic range of the image is from zero (white) to about $5 \times 10^{-4} \text{ s}^{-2}$ (black). Tropopause markers as in Figure 6.10.

biteouts. As a cautionary note, however, one should be aware that the contouring algorithm employed in Figure 6.18 accentuates the appearance of “layers”; although the raw data certainly show some tendency towards layering (in the stratosphere in particular), the strong layering depicted is misleading.

The final field we present is the profiler vertical Doppler beam, hourly-average signal-to-noise ratio (SNR), displayed in Figure 6.19. Recall from Chapter 4 that SNR is proportional to returned power, and that the profiler is sensitive to moisture and temperature gradients; essentially, to (pressure-scaled) atmospheric stability (Equation 2.18). The regions of near-adiabatic lapse rate which have been discussed at length previously in this chapter are clearly visible, with anomalously low SNR (and, by extension, returned power). The tropospheric enhancement is obvious, at least while it is located below about 10 km. The upper tropopause does not show up as

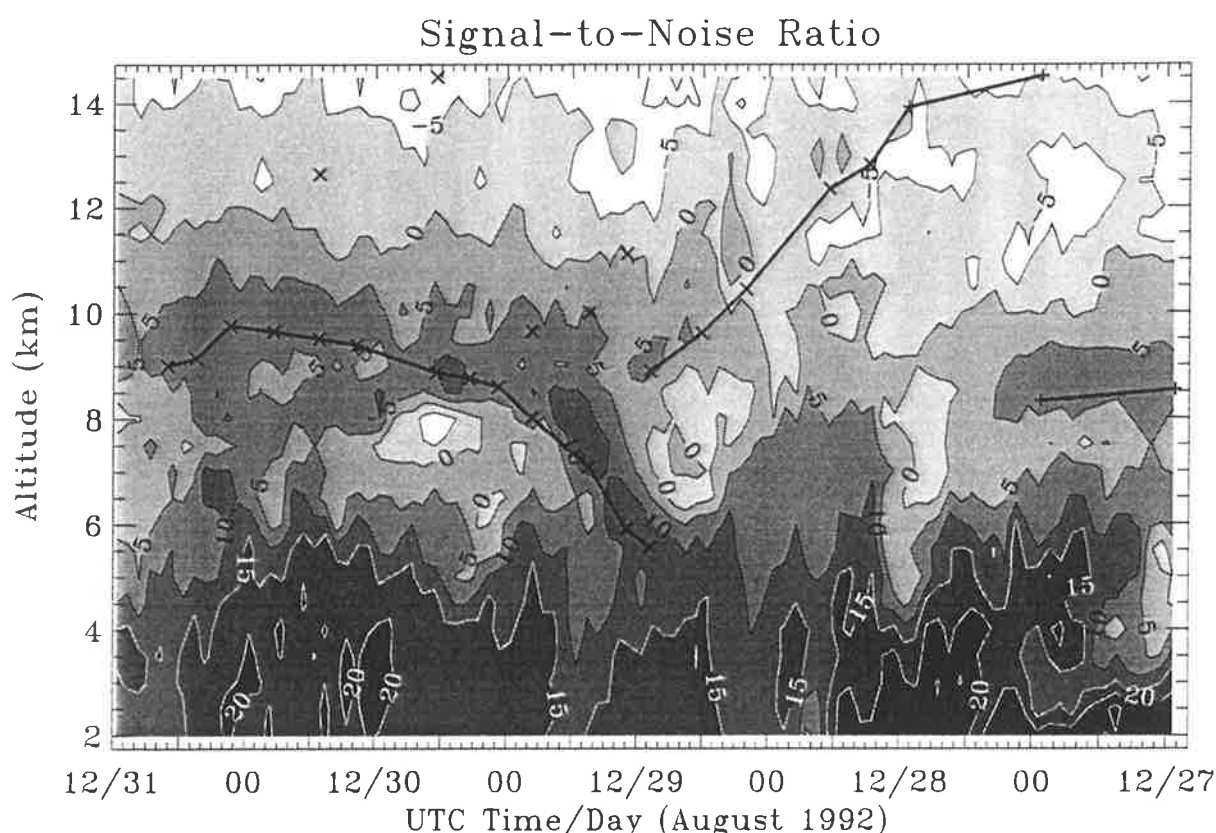


Figure 6.19: Time-height cross-section of radar hourly-average SNR on the vertical Doppler beam, for IOP IV/3; the contour interval is 5 dB. Tropopause markers as in Figure 6.10, with the lower tropopause early in the figure also highlighted.

clearly, although there is a signature between 0300 UTC and 1200 UTC 29 August, and traces as far back as the start of observations; the background exponential decrease with altitude dominates at the upper range gates. However, the profiler clearly picks up the tropopause near 8 km in the first twelve hours of observations (the Adelaide Airport tropopause height determinations have been highlighted). The disappearance of this signal, and the rapid lowering of the upper tropopause in the warm air (which would be far clearer in the returns of a higher-power profiler) are clues for forecasters aiming to predict weather in the short-term. Comparison with Figure 4.4 reveals that the decreased low-level SNRs around 0200 UTC and 1800 UTC 29 August correspond to times of increased spectral width due to turbulent contributions. The effects of moisture can be seen by comparing with Figure 6.9(b), for example: the dry slot causes a reduction in returned power through the mid-levels, while the enhanced moisture due to the subtropical front between 1200 UTC 28 August and about 0300 UTC

29 August results in increased returns through the middle- to upper-troposphere. In conjunction with the measurement of wind profiles, the ability of the radar to observe these changes in atmospheric structure in real-time is a powerful tool for nowcasting efforts.

6.5 Discussion

In this section we investigate the cause of the intense rainfall associated with this event. By 1100 UTC 29 August, the warm tongue and cloud head have entered the centre of the surface low, while the cold tongue and dry slot have advected cyclonically around the western side of the surface low; the leading edge of the cold tongue lies over Adelaide (Figure 6.2(c)). The tropopause height has descended to 5 km altitude, with a cold, potentially unstable airmass lying over Adelaide.

Moderately heavy rain began falling in Adelaide with the arrival of the cloud head at 1530 UTC 29 August; four hours later a narrow convective band of heavy precipitation passed through, with rainfall rates exceeding 40 mm/hr and surface winds gusting to 25 ms^{-1} [Furler, 1993]. The convective band intensified as it crossed the Mount Lofty Ranges (circa 700 m ASL), although Mills & Wu [1995] note that its origin was south of Kangaroo Island some six hours earlier.

Two SAFEX sonde flights from 29 August are shown in Figure 6.20. The dark lines represent the temperature and dewpoint for the 1257 UTC sounding, while the lighter lines show data from the 1628 UTC flight. There are two important points to note about the later flight: it displays significant warming between 850 hPa (1.25 km) and 450 hPa (6 km) and then cooling aloft up to 300 hPa (8.75 km) when compared to the earlier flight; and it is significantly more moist throughout the troposphere, attaining near saturation. Figure 6.21 displays the 1628 UTC flight again, but this time with the 2141 UTC flight. Note how the upper troposphere has dried out significantly over just five hours, while the process of warming below and cooling aloft has continued. This combination of warming, moist air below cooling, dryer air aloft destabilises the atmosphere. By 2300 UTC 29 August, the cold and warm tongues have intertwined, with the cold air confining the warm air to the centre and northern flanks of the low

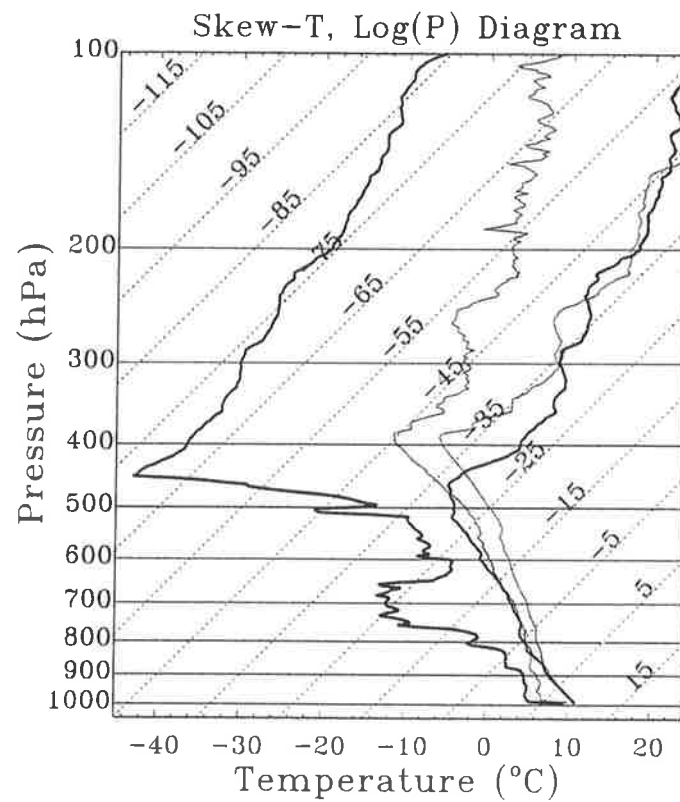


Figure 6.20: Soundings taken at Buckland Park for 1257 UTC (dark) and 1628 UTC (light) on 29 August 1992. Temperature (right-hand profile) and dewpoint (left-hand profile) are plotted for each flight.

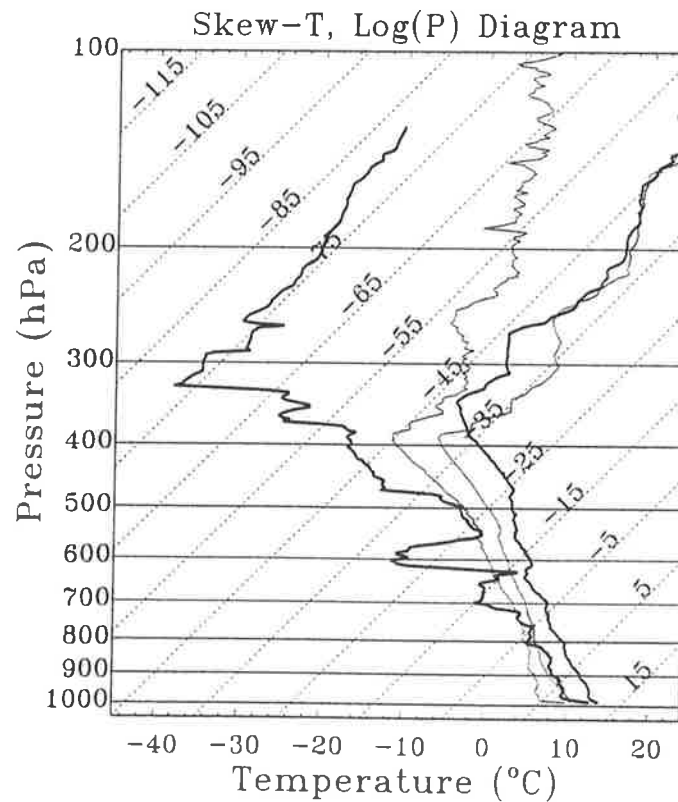


Figure 6.21: As in Figure 6.20, but for flights at 1628 UTC (light) and 2141 UTC (dark) 29 August.

(Figure 6.2(d)). The warm-air advection to the west and northeast of the low which we have discussed previously leads to ascent, accentuated by the flow over the Adelaide Hills. When combined with the potentially unstable situation of cool dry air overlaying warm moist air, the triggering of convection is practically guaranteed.

6.6 Conclusion

We have documented the development and passage of a cut-off low to the south of Australia with a combination of VHF wind profiler data, radiosonde retrievals, automatic weather station time series and RASP analyses, in detail unprecedented in the region. The system was shown to have remarkable similarities to the “instant occlusion” process studied by *Browning & Hill* [1985], featuring a dual conveyor belt pattern. Two separate systems—a subtropical cloud band with associated jet and surface front, and a low-pressure system south of the continent with an associated upper PV anomaly—were shown to have interacted with each other, while retaining their separate identities. As the subtropical front strengthened, the midlatitude temperature gradients weakened. Circulation around the midlatitude system caused cold air to encircle the low and cut off a pool of relatively warm air. The advection of warm, moist air at low levels, and colder, dryer air at upper levels, acted to destabilise the atmosphere, with severe convection triggered as the moist air was lifted further over the Adelaide Hills area. All of the diagnostics we have presented here—soundings, isentropic flows, trajectories, and profiler retrievals—are consistent with the model of potential-instability initiated convection.

Comparison of time-height sections of winds observed by the radar and those observed by the radiosondes are compared and show good agreement. Comparison of these cross-sections with zonal cross-sections from RASP show good broadscale agreement. In particular, the overall pattern of subsidence and ascent (the mean vertical wind) over the domain is duplicated by the RASP analyses and profiler retrievals. Matching the time resolutions of the two data sources significantly improves the quantitative comparison, but differences due to evolution in time and space (especially in

the rapidly-evolving vertical wind field) present a challenge which may well prove intractable. It is interesting to note that RASP, with a grid-size of 150 km, can resolve mesoscale features such as the low-level pool of moist air, and the elevated region of dryer air. Improvements in RASP to a 75 km grid spacing [Mills & Logan, 1994] and more recently to a trial 10 km resolution (P. May, private communication 1996) advance the possibility of even more appropriate comparisons with radar wind profiling systems. The vertical wind is a key component in forecasting meteorology, and direct measurements (such as those provided by the profiler) have the potential to greatly enhance our understanding of mesoscale events such as the case study presented here. The confirmation of profiler vertical wind accuracy via comparison with accepted sources, which has to date been poorly tested, is an essential step towards the widespread acceptance of such measurements.

Chapter 7

THE PREFRONTAL TROUGH

For every problem, there is one solution which is simple, neat and wrong.

— *H. L. Mencken*

7.1 Introduction

The passage of cold fronts is a major feature of synoptic charts covering southern and southeastern Australia during spring and summer, bringing cooler temperatures and occasionally precipitation. The summer fronts are generally dry; precipitation (and moist processes in general) serve to complicate many of the spring events. These fronts are quite different from their counterparts in Western Europe and the United States, with development over the ocean to the south accentuated by a large land-sea temperature gradient at the predominantly east-west coastline. The general synoptic situation sees a cold front associated with a poleward parent cyclone (well south of the continent), with frontogenesis occurring in the region between two subtropical anticyclones. The Cold Fronts Research Programme [*Smith et al.*, 1982; *Ryan*, 1982; *Ryan et al.*, 1985; *Garratt*, 1988] investigated these mesoscale events in some detail, including the development of a conceptual model of the air flows associated with them [*Ryan & Wilson*, 1985; *Wilson & Stern*, 1985]. The main feature of this model are: ascent in the warm continental northerlies ahead of the front; subsidence in the cool maritime air behind; a frontal transition zone (FTZ) between, consisting of (possibly several) change lines; and a cold dome near 500 hPa ahead of the surface

front, presumably due to convection in the FTZ. Further information can be found in the comprehensive review of *Reeder & Smith* [1992].

Our interest here is in a poorly-studied subset of these fronts; continuing the theme of Chapter 6, the interaction of a surface cold front (SCF) with an associated prefrontal trough (PFT). *Hanstrum et al.* [1990a] developed a climatology of such events, where troughs migrating ahead of an approaching midlatitude Southern Ocean cold front can produce a significant cool change in their own right, due to the change in air trajectory around the trough; low-level convergence along the trough can also lead to convection. Together with a companion case study (*Hanstrum et al.* [1990b]), these are the only studies devoted to the PFT which we are aware of, although there are a few case studies and analyses which do not specifically consider the details of the PFT despite its presence (see, for example, *Garratt* [1984], *Physick et al.* [1985], *Garratt* [1988], *Ryan et al.* [1989] and *Tapp & Oliver* [1989]).

During the warmer months (September to April), surface troughs regularly form in central Western Australia (see Figure 6.1 for locations), primarily due to surface heating [*Fandry & Leslie*, 1984]. As a Southern Ocean cold front approaches, this “prefrontal” trough translates eastwards (Figure 7.1, later in this chapter, shows a typical sequence). At times, rapid frontogenesis can occur within the PFT, usually at the expense of the original SCF: compare the situation from Chapter 6, where the strengthening of the subtropical front at the expense of the midlatitude system was a significant feature. The analysis and prediction of this interaction is a major problem for local forecasting efforts, and the lack of investigation into it is surprising. In particular, the underestimation of frontal speed is a major source of forecast error [*Wilson et al.*, 1987], and an observed “acceleration” of the front is often associated with the presence of a developing PFT. It is partly due to difficulties in identifying the analysis sequence associated with development that acceleration is assumed, but the lack of a widely accepted model of frontogenesis in the PFT accentuates the problem. Hence, the forecaster is uncertain as to the degree of development to attribute to the trough, especially early in the sequence. There appears to be an inclination towards underestimation (with the forecaster tending to trust the originally analysed SCF ahead of development in the PFT), resulting in an apparent discontinuity in the movement of

the “front”—the cool change arrives significantly earlier than predicted. The statistics presented by *Hanstrum et al.* [1990a] are illuminating: by the end of the development sequence, the original front was present and identifiable in only 15% of cases, while the PFT was analysed as a cold front in 50% of cases; both SCF and PFT were still present as separate features in only 3% of their ten-year, 150-case dataset.

Some of this confusion arises from differences in the concept of a “front” under different theories. There are three main schools of thought, which have been reviewed by *Smith & Reeder* [1988]: air mass theory, which postulates that a front is a narrow transition zone between two air masses; the gravity current approach, where the frontal speed is approximated by formulae appropriate to laboratory gravity currents (which can have some application to real cases, but the errors involved in physical measurement of the required parameters in the real world lead to rather wide bounds on the eventual estimate); and frontogenetic models, which we shall briefly discuss.

Assorted deficiencies in the air mass and gravity current concepts (generally related to differences between observations and predictions) sparked interest in a more mathematical approach. The modern theory of adiabatic inviscid frontogenesis has been reviewed by *Hoskins* [1982]: to summarise, frontogenesis is initiated by a deformation in the synoptic wind field, which advects parcels relative to each other, leading to the development of temperature and wind gradients. Changes in the horizontal temperature gradients lead to vertical circulation, which enhances convergence in the warm air ahead of the front and results in an intensification of the front. However, the ascent of warm air produces adiabatic cooling (similarly, subsidence leads to warming), which acts to counter frontogenesis. Hence, development and intensification are only seen near the surface, or near the tropopause, where vertical velocities are suppressed.

Under this framework, we would only expect to see fronts in regions of active frontogenesis, since turbulent diffusion acts to weaken gradients in the absence of such development. Frontogenetic approaches suggest that a front is *not* a quasi-stationary, impermeable boundary between distinct air masses; rather, it is a region of maintained synoptic-scale frontogenesis. There is considerable modelling and experimental justification for this position, as described by *Smith & Reeder* [1988] and later (with more direct application to southern Australian fronts, and a discussion of the experimental

observations of *May et al.* [1990]) by *Reeder & Smith* [1992]. If the front were material, for example, then the surface would coincide with a zero relative-flow isotach; however, both theoretical models and observational evidence points to a low-level cross-frontal flow, from the warm air to the cool side. In many cases of observed cold fronts, the “front” moves faster than the normal component of wind at any level, ahead or behind the front. Hence, the front *propagates* relative to the warm and cold air masses—it translates by development rather than by advection, with a speed linked to that of the baroclinic wave in which it is embedded.

Reeder & Smith [1992] present an initial theory of the development of the PFT based on frontogenetic arguments. As the SCF approaches Western Australia, the combination of the wind field around the parent low-pressure system, the surrounding anticyclones, and the northern heat low/trough system over the Australian continent, result in a pattern of synoptic-scale deformation. This enhances the temperature gradient between ocean and land, and promotes frontogenesis in the region of the PFT. With the front translating by development rather than advection, it is relocated towards the PFT. If we define the location of the front at the surface by the drop in surface temperature, then baroclinic wave models (see, for example, *Reeder et al.* [1991]) predict that the front lies to the west of the trough axis, with the separation decreasing as the system intensifies.

In the only detailed case study of the prefrontal trough to date, *Hanstrum et al.* [1990b] found that development in the trough led to strong winds backing by 90° or more, a significant drop in temperature, and the onset of cloud—a picture consistent with the passage of a cold front at the surface. The system described briefly by *Garratt* [1984], and in more detail by *Garratt* [1988], associates the PFT with a shallow region of cool air, vigorous wind shift (backing) with gusts at the leading edge, large surface temperature gradient, and a pressure jump; the originally analysed SCF, on the other hand, was still present on the forecasting charts, although perhaps more for reasons of continuity with the previous analyses. The system analysed by *Tapp & Oliver* [1989] features a prefrontal trough where, at the time of passage over the continent, “the main changes in weather elements over southeast Australia were associated with what *had been* the prefrontal trough” (our emphasis). These features are all associated

by the investigators with “rapid frontogenesis in the trough”, leading to an apparent discontinuity in the surface position of the change over a period of twelve hours or so. Generally, the SCF is analysed as present until after the passage of the change at the surface, at which point it is combined with the PFT as a single system. As mentioned above, the analysis of the SCF is often for simple continuity, or by association with a cloud band visible in satellite imagery (presumed to be associated with ascent along the line of the SCF).

However, this identification on the basis of cloud may well be inappropriate. *Tapp & Oliver* [1989] consider the development of mid-level cloud ahead of the SCF (this development frequently being associated with a PFT) and the evolution of the frontal cloud mass. This is an important forecasting problem, since in the absence of cloud there will be strong solar heating leading to a large temperature differential between continental and maritime air, while cloud well ahead of a front will suppress such warming (as indeed will the evaporative cooling of precipitation). The presence of cloud is affected by two factors: available moisture (q), and the degree of ascent (w). If q and w do not change, then any cloud will remain in equilibrium between ascent, entrainment and precipitation; if they subsequently change, then there will be a change in the degree of cloudiness.

Tapp & Oliver [1989] model the distribution of vertical motion through the quasi-geostrophic omega equation (Equation 3.4) for a number of cases, and find a delay of 12–24 hours between changes in vertical ascent and resultant changes in the cloud field. Extensive cloud development in their study is associated with regions of widespread ascent, particularly along a south-eastward path stretching from the central Australian continent towards the Bight (the warm conveyor belt of the conceptual cold front model presented in *Ryan & Wilson* [1985]). Of course, moisture is still a critical component—ascend in dry air will not form cloud, while weak ascent in a very moist atmosphere will. However, ascent is certainly a required component for the formation of cloud; once cloud has formed, the onset of localised subsidence leads to its eventual erosion. *Tapp & Oliver* [1989] note that the changeover from ascent to subsidence coincides with the rear of the cloud associated with the frontal zone. We note also the work of *Mills* [1989], who found explosive cloud development following increased

ascent associated with frontogenesis.

Hence, the mere presence of cloud—even organised cloud bands—in a given satellite image is no guarantee of ongoing ascent or active frontogenesis in that region. There are periods of significant length on the mesoscale (and even the synoptic scale) where widespread cloud due to ascent in the past—“fossilised” cloud, perhaps—can exist and advect. Under the right conditions, this advection could occur so as to keep the cloud associated with the propagating region of ascent around the front, prolonging its lifetime as a coherent structure even if it is not an indicator of direct frontal ascent. Similarly, as the front propagates, it will not necessarily generate cloud as it moves: the development of cloud can lag the moving region of ascent, as outlined above. This picture is consistent with the idea of the “front” propagating from the region of the SCF towards the PFT, and is far more satisfying than the forecasting model of a single front undergoing acceleration so rapid as to appear discontinuous from one analysis to another. If the changes in weather at the location of the PFT as it moves over the continent are the same as those associated with a cold frontal passage, the application of Occam’s Razor to the forecast may well be warranted.

The implication is that forecasters should pay attention to development in the PFT when considering the timing of weather changes at the surface, rather than concentrating on the analysed SCF. In particular, the tendency of local forecasters to “invent” perturbations in the surface pressure at the analysed SCF in an effort to justify the analysis of such a feature is regrettable. However, given the paucity of detailed case studies of the prefrontal trough, making precise deductions as to the nature of the PFT evolution is difficult. The importance of the SCF–PFT system to regional forecasting is without doubt: *Hanstrum et al.* [1990a] estimates two occurrences per month outside of winter, which is by no means infrequent; and the large changes in wind direction and magnitude which are associated with the passage of the PFT can have major impacts during the summer bushfire season in Australia, where accurate forecasting is essential in the control of firefighting efforts. With regards to the latter point, the Ash Wednesday fires (16 February 1983) were affected by the passage of just such an event, as detailed by *Garratt* [1984], *Garratt* [1988], and *Reeder & Smith* [1992]. This is a particularly illuminating example, with limited cloud associated with the line of

the “cold surge” (PFT) cool change, but significant cloud around the analysed SCF (which did not display a temperature, wind, or significant pressure change).

We now examine a prefrontal trough case study from the SAFEx programme, IOP IV/7.

7.2 Discussion

From the outset, it is worth noting that single-station profiler and sonde retrievals are not well-suited to studying this phenomenon, since the area of development is well to the south and west, over the ocean. An examination of horizontal cross-sections would be a valuable addition, which should be considered in future case studies. However, the available data is exceptionally good for nowcasting the change, and gives a comprehensive picture of the system as it passes just inland of the coast. Additionally, the use of serial soundings to represent a time (or distance) versus height cross-section assumes steady-state behaviour; in this sort of situation, with an evolving system (active, propagating frontogenesis), the steady-state assumption is invalidated, but the analysis is still qualitatively useful (especially since the major development has already occurred by the time the system reaches the observing site).

Figure 7.1 shows a sequence of four MSLP analyses from the Bureau of Meteorology. In frame (a), 48 hours before passage, Adelaide is under the influence of a high-pressure system, experiencing onshore winds. A surface cold front is approaching from the southwest, while a trough is present over the western coast of the Australian continent. Frame (b), a day later, shows that the trough has begun moving across the Bight, while the SCF is analysed some distance behind. Note how the isobars near the SCF have been drawn with slight “cusps”, as the forecaster follows conventional wisdom (in the absence of actual surface measurements) regarding wind changes at the analysed SCF. The corresponding satellite image is displayed as the top panel of Figure 7.2, where the cloud band associated with the analysed SCF is the dominant feature. Note that the prefrontal trough has already developed a significant cloud band of its own, and that the air to the rear of the SCF-cloud is characteristic of ascending cold air (the “speckled” pattern); that is, there is no strong subsidence behind the

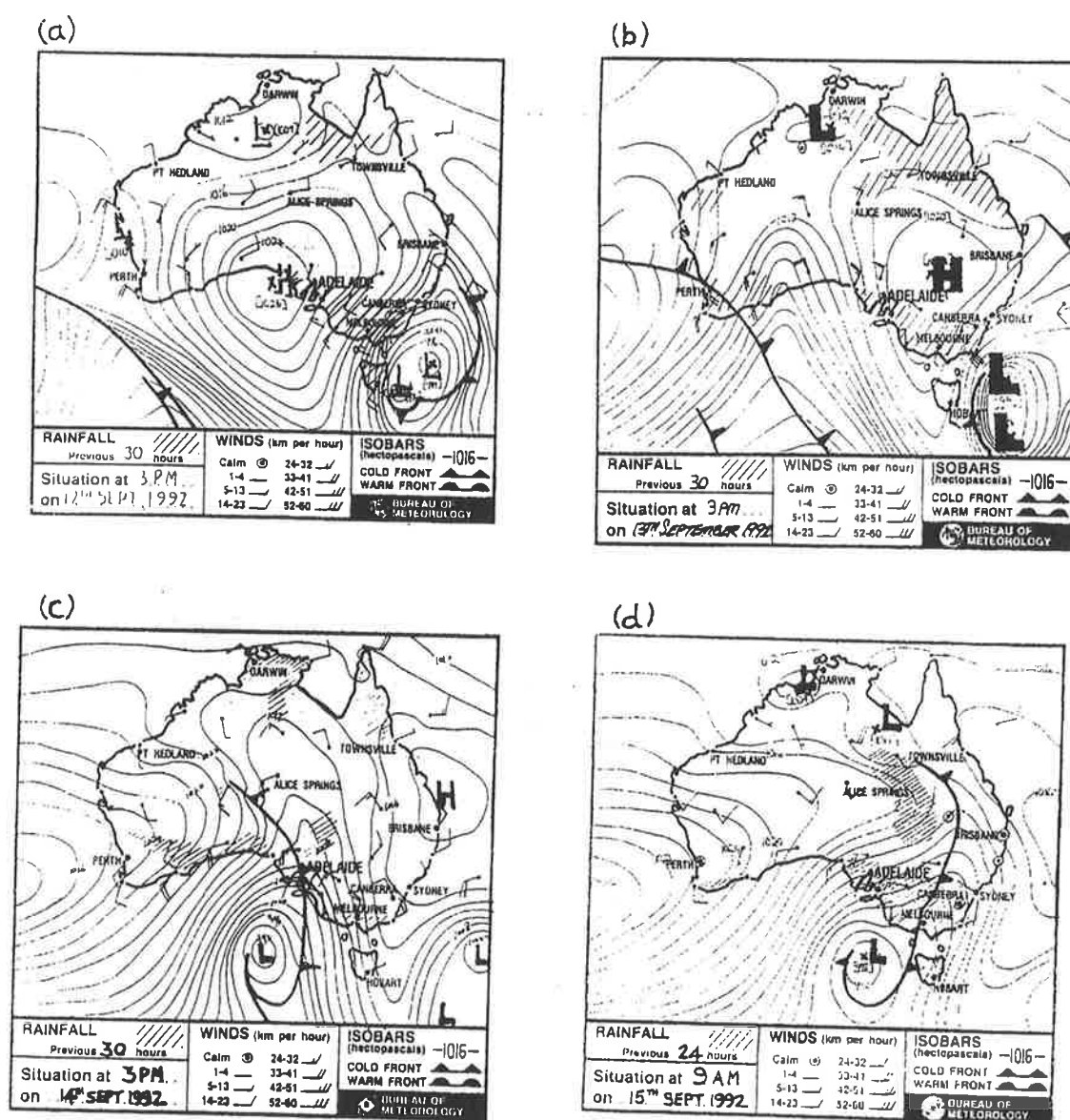


Figure 7.1: MSLP during IOP IV/7 (1992): (a) 0600 UTC 12 September; (b) 0600 UTC 13 September; (c) 0600 UTC 14 September; (d) 0000 UTC 15 September.

analysed SCF even at this early stage. Figure 7.1(c) is closest to the passage of the event over Adelaide, and the SCF is being analysed purely on the basis of the cloud band seen in the lower panel of Figure 7.2; the PFT cloud band is now dominant. Note also the thin band of cloud extending across Western Australia and into South Australia, reminiscent of the North-west cloud band of Chapter 6, and presumably (given our lack of other synoptic fields) associated with the warm conveyor belt as

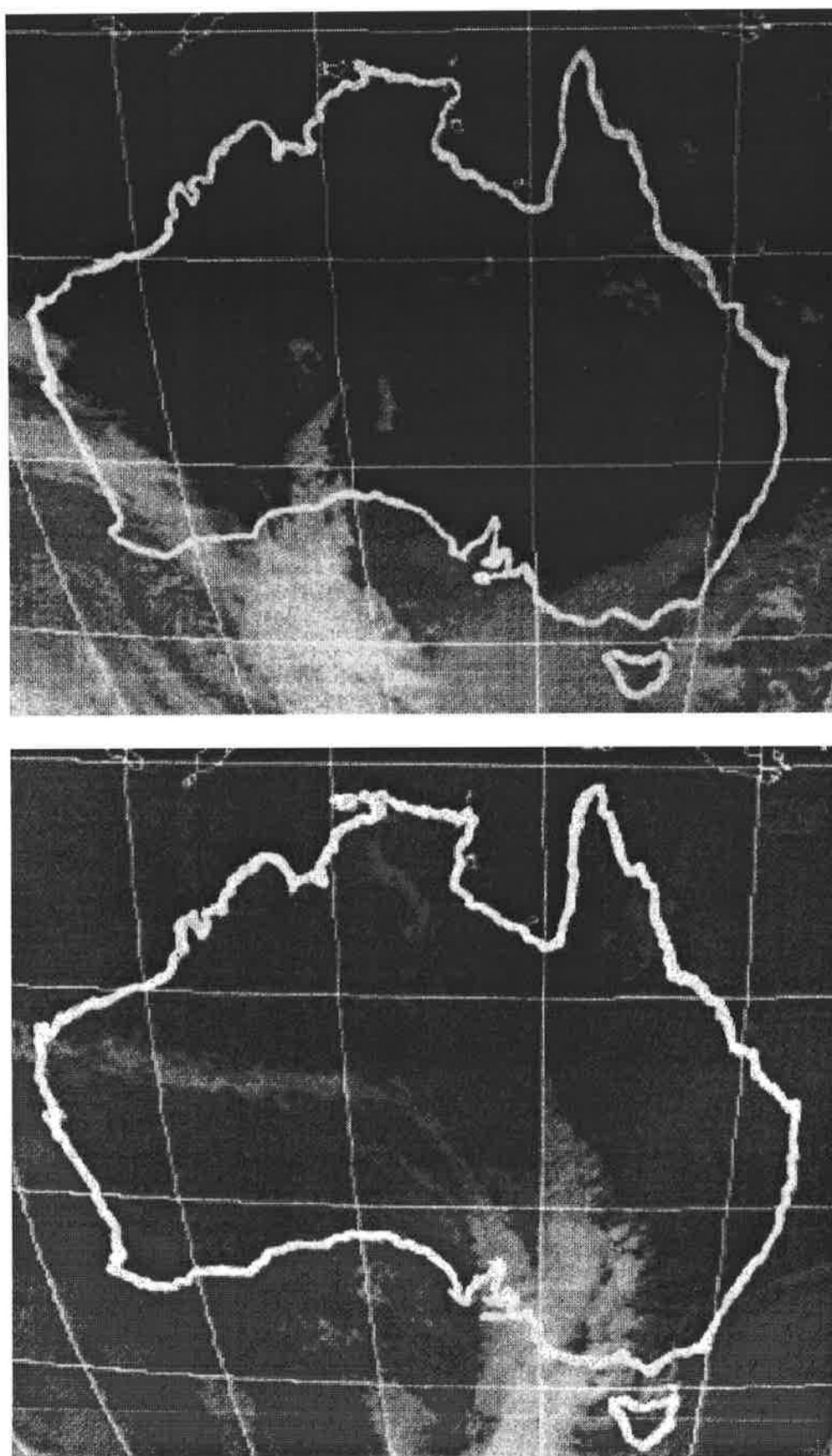


Figure 7.2: GMS IR satellite images from 0300 UTC 13 September 1992 (top) and 0300 UTC 14 September 1992 (bottom).

described earlier. The SCF cloud band is still followed by speckled cumulus (cold air ascent). The final panel (d) of Figure 7.1 shows the SCF and PFT now analysed as the same feature; the relevant satellite image (not shown) comprises a single cloud band over the east of the continent.

The main local effects of the passage of the system centered on flooding of the Gawler River (two kilometres south of the Buckland Park field site) on 15 and 16 September, which inundated several hundred hectares of agricultural land in the surrounding Virginia and Two Wells townships.

Automatic weather station data from Buckland Park is shown in Figure 7.3, analysed in the same way as the corresponding data presented in Chapter 6. The minimum surface pressure (panel (a)) occurs at 0500 UTC 14 September, and corresponds well with the temperature change (although there is some masking from the diurnal variation, since the passage time is about 1430 LST). However, the 5°C warming from 1230 UTC (2200 LST) 13 September to around 1900 UTC 13 September (0430 LST 14 September) is certainly against the diurnal trend! Note that this front, occurring in mid-September, is not expected to be associated with the same degree of temperature change experienced in summer fronts. The wind change (panels (b) and (d)) is consistent with a single change line at around 0500 UTC 14 September, with associated wind gusts (panel (c)). The timing of this change line coincides with the passage of the PFT, using the speeds of the analysed features as marked on Figure 7.1 (see panel (c) of that figure, for example, where the SCF is still some distance offshore). We find the conventional explanation, which would be to invoke a rapid acceleration or dislocation of the SCF towards the PFT, to be unnecessary, if not unreasonable.

The cross-section of potential temperature derived from seventeen Marwin flights over sixty hours is displayed in Figure 7.4(a), together with significant relative humidity contours. Note the strong similarities with the IOP IV/3 situation (Figure 6.10): the tropopause is raised in the warm air ahead of the front and lowered in the cool air, with an exchange of potential isotherms between troposphere and stratosphere through the tropopause “break” region near 7 km altitude at 1200 UTC 14 September. Again, note how the isentropes bend toward the tropopause perturbation from above and below. There is a pronounced dry slot near 0700 UTC 14 September. Equivalent potential

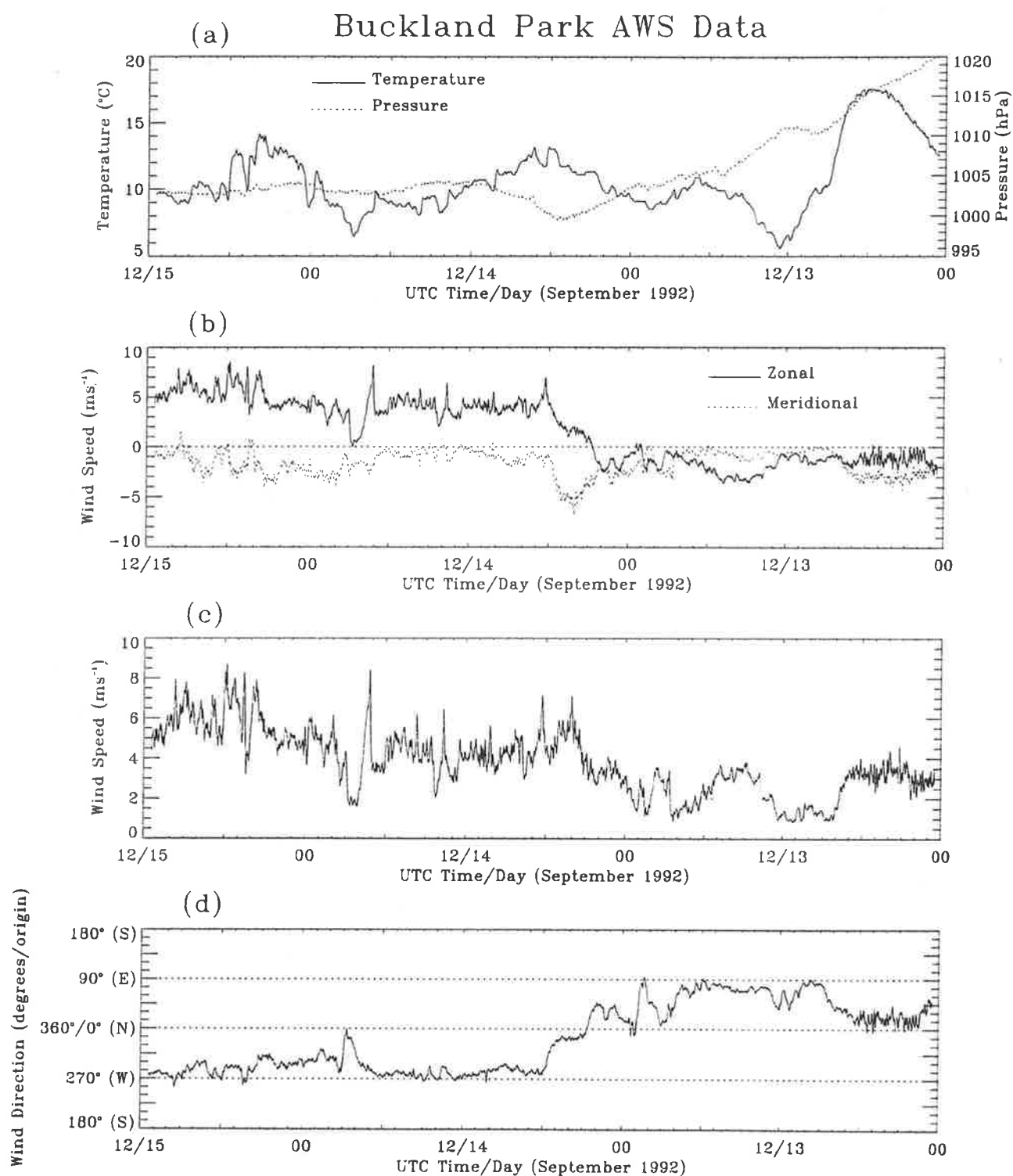


Figure 7.3: Buckland Park automatic weather station data during IOP IV/7: (a) temperature (solid, $^{\circ}\text{C}$) and pressure (dotted, hPa); (b) zonal (solid) and meridional (dotted) wind components (ms^{-1}); (c) wind speed (ms^{-1}); and (d) direction of origin of wind.

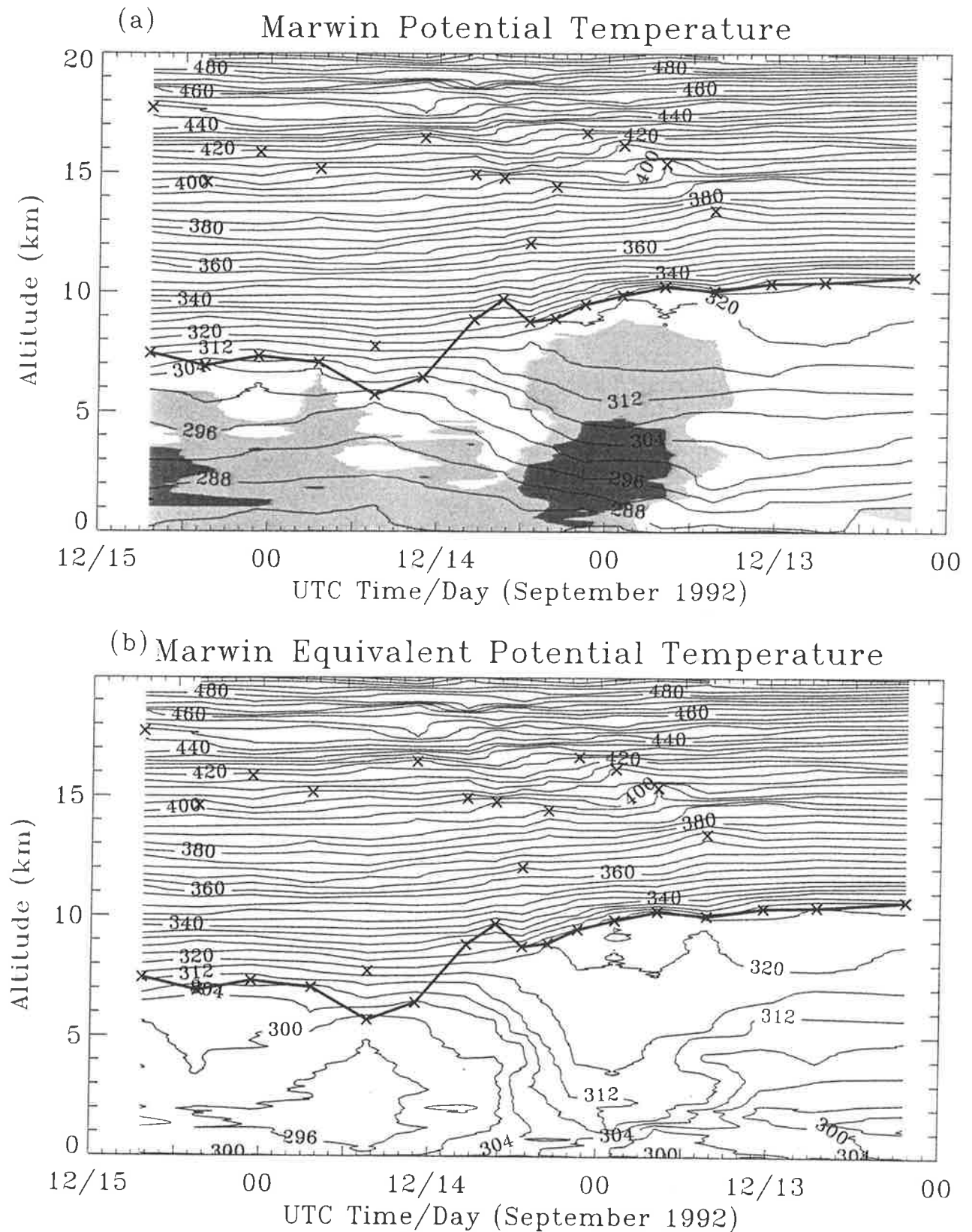


Figure 7.4: Time-height cross-sections for IOP IV/7 from SAFEx radiosonde data: (a) potential temperature (solid lines, 4 K intervals) and relative humidity (shaded at 60% and 90%); and (b) equivalent potential temperature (4 K intervals). The sonde-determined tropopause heights are marked by $\times$'s, with the major lower-tropopause estimates joined by a heavy solid line.

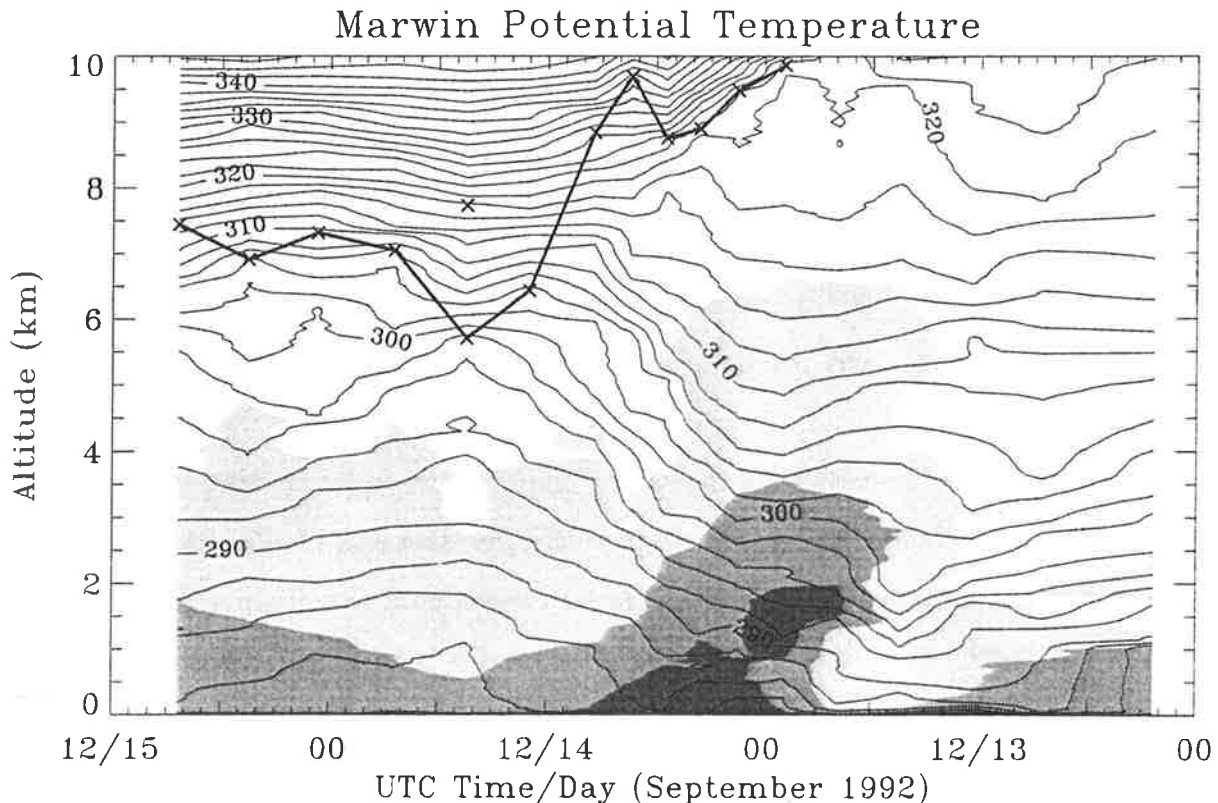


Figure 7.5: Time-height cross-sections for IOP IV/7 from SAFEx radiosonde data of tropospheric potential temperature (solid lines, 2 K intervals) and mixing ratio (shaded at 1, 4 and 6 g kg⁻¹, lightest to darkest). Tropopause markers as for Figure 7.4.

temperature is contoured in Figure 7.4(b), with well-mixed cold air a feature of the post-frontal situation (near 1800 UTC 14 September), while the warm prefrontal air is marked by weakly-stable conditions at middle-levels around 0000 UTC 14 September. Generally, the upper front is quite strong, while the surface front is not so well marked.

Figure 7.5 concentrates on the lowest ten kilometres, and uses mixing ratio as a moisture tracer. The low-level structure is much more apparent, especially around 0500 UTC 14 September where the temperature gradient into the cold air is most marked. The cold dome is somewhat further into the warm air than in a “classic” front (see *May et al.* [1990], for example), and slopes forwards with height, but this is likely a reflection in the single-station measurements of the intense system development. Warming through the lowest few kilometres centered around 1600 UTC 13 September (0230 LST 14 September) is consistent with the AWS surface measurements, and the available moisture matches with the cloud imagery. There are many similarities with

what *Browning* [1990] refers to as a “warm conveyor belt with forward-sloping ascent” (see his Figure 8.10(b) in particular), where an upper θ_e front runs ahead of a surface cold front: in particular, the warm, dry air under-running elevated moisture (1500 UTC to 2100 UTC 13 September, under 2 km altitude); and cold, dry air overrunning a shallow moist zone behind the upper front (after 0900 UTC 14 September, above 1 km). *Browning* [1990] notes that this “split” cold front is common in the United Kingdom, and regularly results in forecasting confusion when simple classical frontal models are applied.

Marwin wind retrievals are shown in Figure 7.6: the most striking features are the zonal and meridional jets located, respectively, just above and below the tropopause in the warm prefrontal air. The contour of zero meridional velocity, marking the change from northerly to southerly flow, slopes backwards with height towards the lowered tropopause, and intersects ground level at about 0500 UTC 14 September (consistent with the analysis above). The cyclonic structure is clear, with particularly strong cyclonic shear in the region of the tropopause break.

The profiler was operated in purely vertical Doppler mode through IOP IV/7, which in many ways is unfortunate: although we have excellent resolution in the vertical velocities, we have no profiler time series of the horizontal wind. The profiler data is also interspersed with four data gaps, each of an hour or two duration, due to another experiment competing for observing time. Figure 7.7 shows the SNR on the vertical Doppler beam throughout the IOP, with the tropopause break showing up clearly. Note that we have chosen to display this figure on the same vertical scale as the Marwin plots, to give a comparative picture of the behaviour in the stratosphere—in particular, the upper-tropopausal enhancement near 15 km altitude. Weakly stable regions in the troposphere are again seen to correspond with areas of reduced SNR; the passage of the front is preceded by reduced low-level SNRs for about twelve hours, while prior to that the low-level SNRs were particularly strong (compare with Figure 6.19, which shows precisely the same behaviour).

Profiler vertical velocities are shown in Figure 7.8, and have been corrected for beam-tilt contamination using the Marwin horizontal wind fields (interpolated via a cubic spline routine to a matching time resolution of one hour). The region of

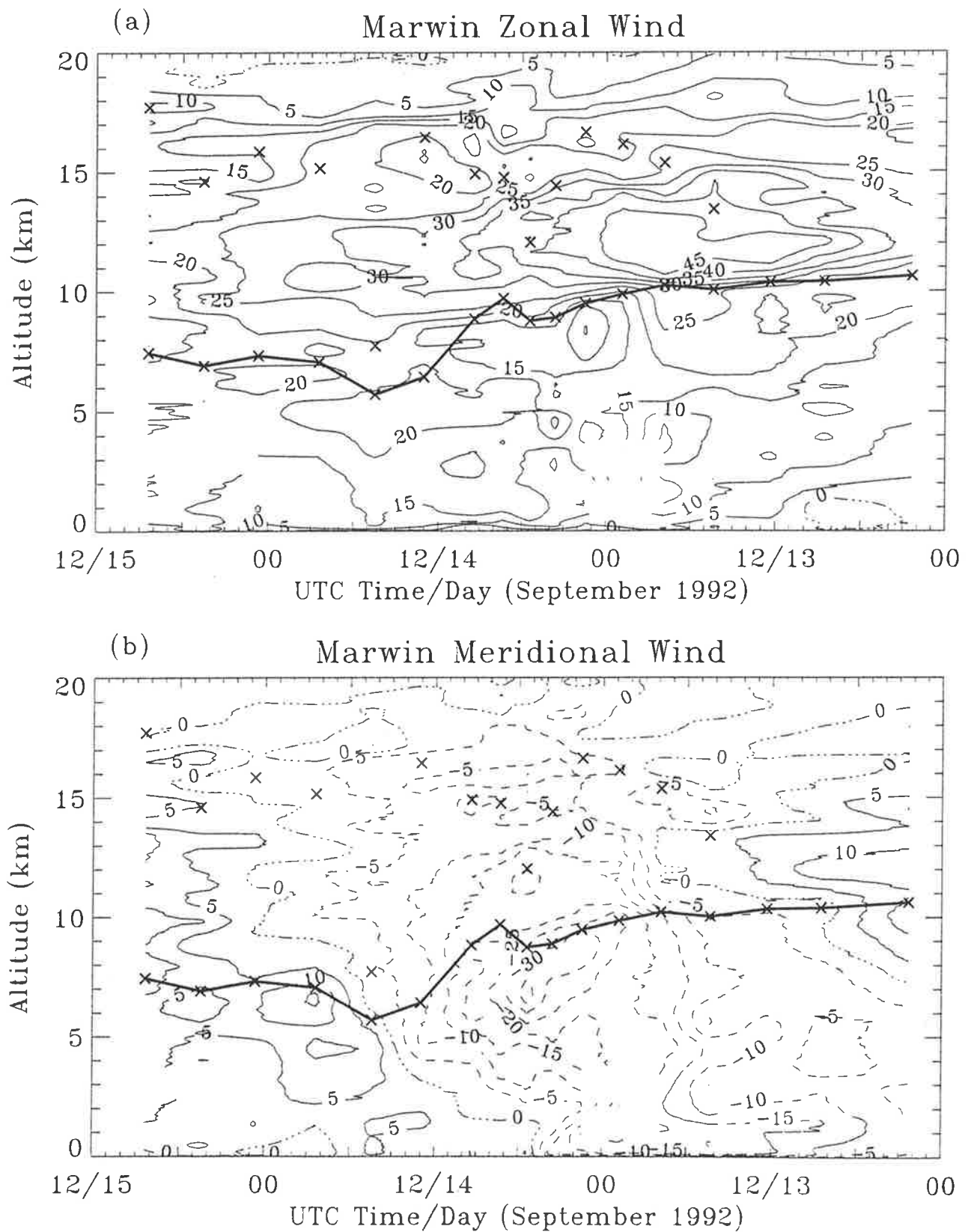


Figure 7.6: Time-height cross-sections for IOP IV/7 from SAFEx radiosonde data: (a) zonal wind; and (b) meridional wind (5 ms^{-1} intervals, positive values solid, zero-contour triple-chained, negative values dashed). Tropopause markers as for Figure 7.4.

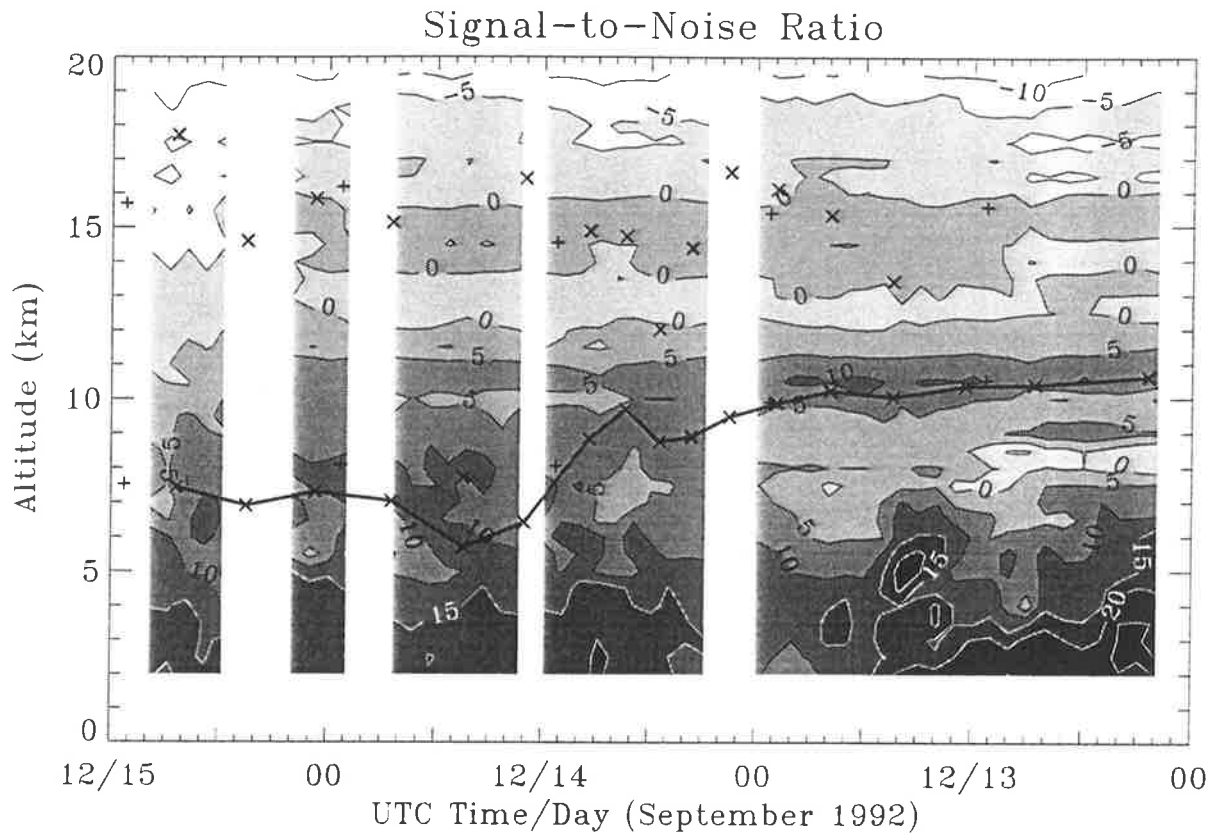


Figure 7.7: Time-height cross-section of radar hourly-average SNR on the vertical Doppler beam, for IOP IV/7; the contour interval is 5 dB. Tropopause markers as for Figure 7.4, with the addition of Adelaide Airport sonde estimates (marked by +).

missing data (and apparent subsidence) around 2200 UTC 13 September is associated with contamination due to precipitation producing large downward velocities; the remainder of the picture reflects air motions. The same general pattern as observed in Chapter 6 is present: mid-level ascent in the moist warm conveyor belt between 2000 UTC 13 September and 0400 UTC 14 September (compare with Figure 7.4(a) for the distribution of moisture); subsidence under the lowering tropopause (to 1000 UTC 14 September); then low-level ascent in the cool air. These regions of vertical velocity phase slope backwards with height, and are extremely well matched to the satellite imagery and proposed model of the SCF-PFT system. In detail, the mid-level ascent corresponds to the PFT cloud band (warm, moist air) with the rear of the cloud at the changeover between ascent and subsidence near 0400 UTC 14 September; the region of subsidence is the dry, cloud-free area between the observed cloud bands. This is in turn followed by low-level ascent in cold air, producing the characteristic speckled cumulus

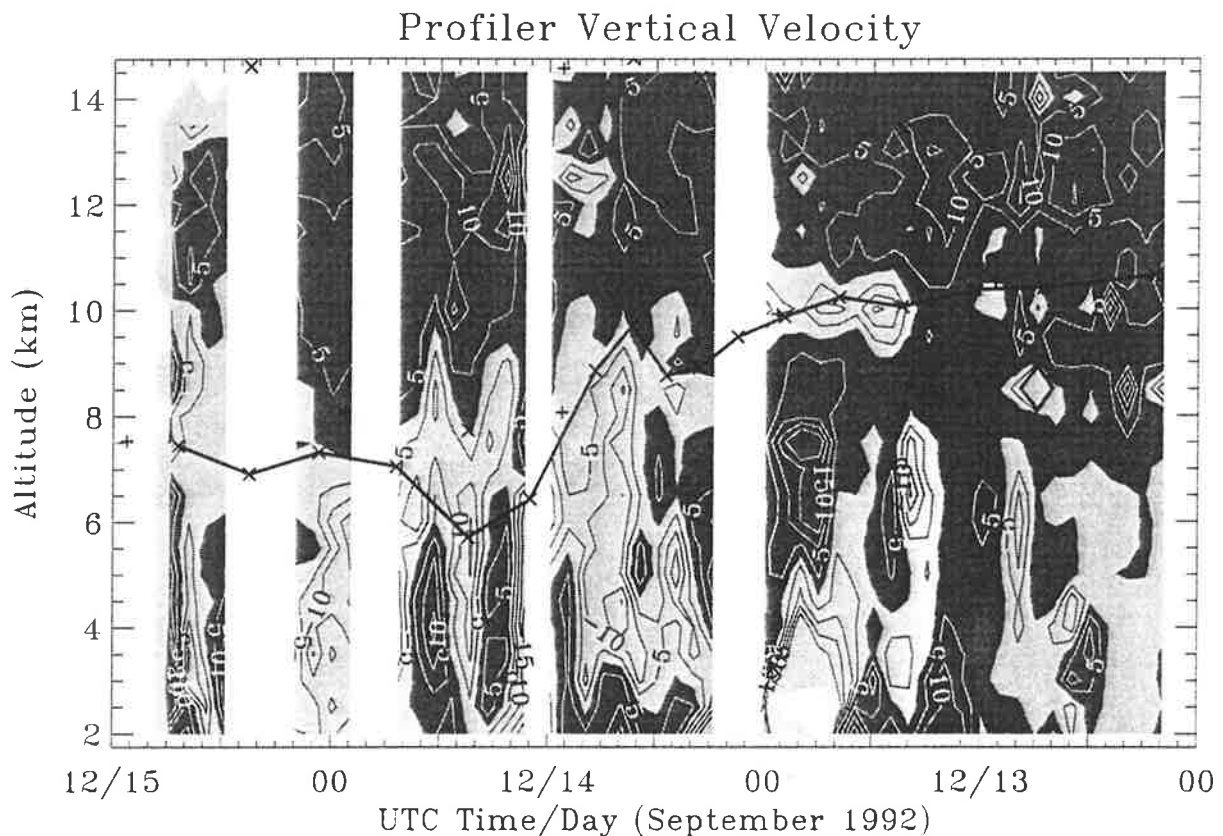


Figure 7.8: Time-height cross-section of profiler one-hour average Doppler vertical velocity (5 cm s^{-1} contour intervals, regions of ascent shaded darker): retrievals corrected for contamination by the horizontal wind due to beam tilt and phasing (see Section 3.1.2). Tropopause markers as for Figure 7.7.

seen in Figure 7.2. The backwards slope with height is also consistent with the concept of an active anabatic cold front, with precipitation (the region of contaminated profiler vertical velocity estimates) at the lowermost extent.

Finally, the vertical velocity activity for the IOP is displayed in Figure 7.9. Ignoring the precipitation-contaminated region near 2200 UTC 13 September, the pattern of activity is basically one of enhanced tropospheric variance around the system, and the penetration of this enhancement into the stratosphere at the tropopause break around 1400 UTC 14 September. This is similar to the IOP IV/3 case shown in Figure 4.2, although the stratospheric penetration there occurs somewhat earlier in the sequence due to the unique nature of the system in question. Note that the tropopause marks a general reduction in vertical velocity activity, as expected.

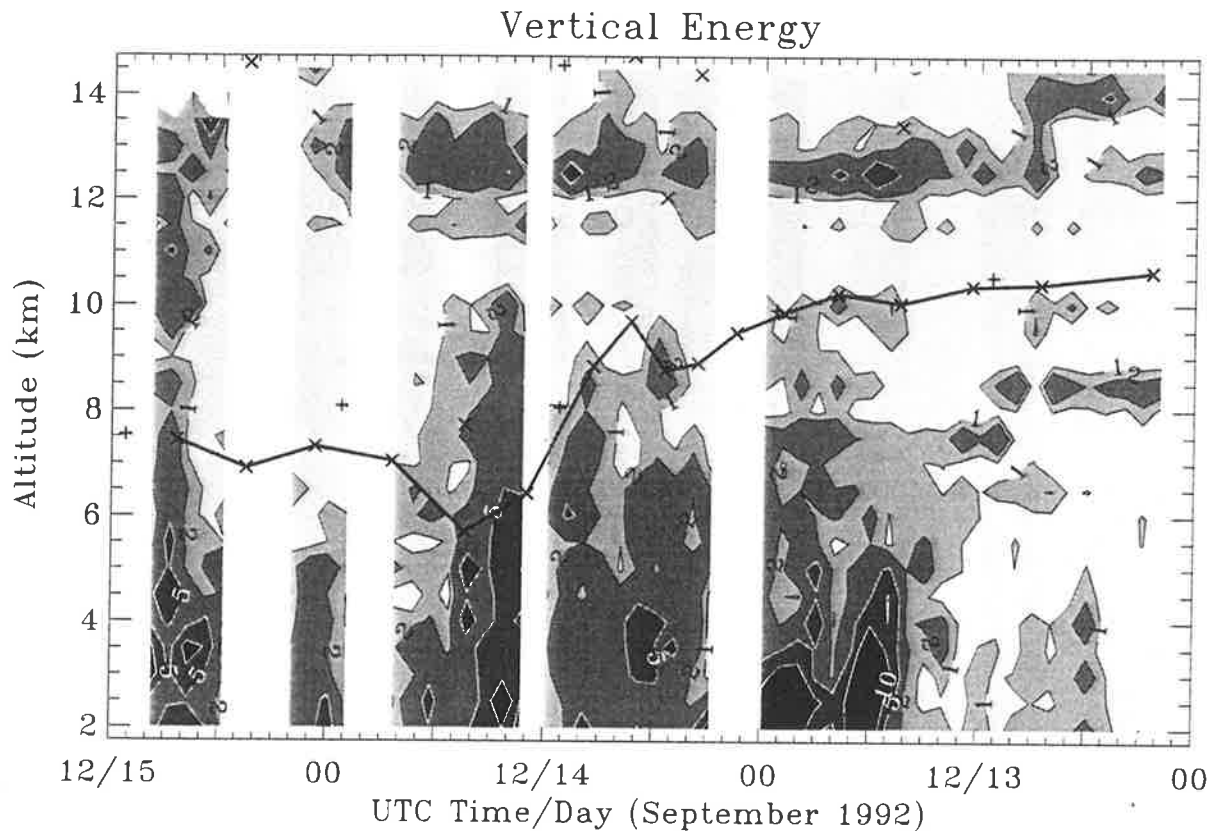


Figure 7.9: IOP IV/7 vertical velocity variance, or energy per unit mass, $\langle w^2 \rangle$, presented as a one-hour average. Contours are non-linear, with units $\times 10^{-2} \text{m}^2 \text{s}^{-2}$. Tropopause markers as for Figure 7.7.

7.3 Conclusions

We have reviewed the meteorology of the prefrontal trough in some detail, and suggested that the frontogenetic picture of development is the most appropriate. A case study from the SAFEx campaign was seen to be consistent with this interpretation: despite the initial analysis of a surface cold front some distance from the prefrontal trough, the pattern of wind, temperature and pressure changes later in the sequence was consistent with a single change line at or near the location of the trough. Often, the remnants of an original SCF can be identified from cloud satellite imagery, but it can be a mistake to associate fronts with clouds in the absence of other convincing tracers. A cloud band can be formed during ascent (frontal or otherwise), and when the propagation (*not* advection) of the original front removes it from the vicinity, the cloud band is left advecting for some moderate amount of time. This is usually of the

order of a day or so, but can be even longer if the cloud band remains associated with the rear edge of the frontal ascent, or with another ascending cell.

Our case study is somewhat limited by the lack of horizontal surface data. *Reeder & Smith* [1988] and *Reeder et al.* [1991] have shown that the axis of maximum relative vorticity provides a robust indicator of frontal position even in the presence of strong surface heating, and this would be a valuable field to include in future analyses tracking frontal propagation between the initial SCF/cloud band and the PFT. It would be even more instructive to examine prefrontal trough events with a full suite of measurements (similar to the case study in Chapter 6), in an effort to tie down the important contributing factors in development to the south and west of the continent. However, the wind profiler's direct measurements of atmospheric vertical velocity are a significant new contribution to the study of these events. Profiler measurements allow for the nowcasting of the motion of features across a particular site, particularly with regards to cloud development (bearing in mind that while ascent is a necessary condition for cloud formation, it is not sufficient in the absence of available moisture). Real-time profiler measurements can be used to give an indication of the slope with height of the changeover between regions of ascent and subsidence, as related to the frontal zone.

Can the prefrontal trough be treated as a front in its own right? Can the surface cold front be assumed to exist throughout the development? Our contention (after *Reeder & Smith* [1992]) is that the actual "front" is often distinct from both features, propagating from the SCF towards the PFT by a process of relocation under frontogenesis. Our observations have borne this viewpoint out for at least one case, although further investigation is essential. As far as forecasting goes, the PFT appears to be a far more appropriate indicator for timing purposes, in the absence of a frontogenetic model of frontal speed. In many ways, this assignment is a fair approximation: track the PFT; observe consistency between the wind, temperature and pressure fields as far as conventional frontal thinking goes; and associate weather changes at the surface with the characteristics of the prefrontal trough.

They hunted till darkness came on, but they found
Not a button, or feather, or mark,
By which they could tell that they stood on the ground
Where the Baker had met with the Snark.

In the midst of the word he was trying to say,
In the midst of his laughter and glee,
He had softly and suddenly vanished away —
For the Snark *was* a Boojum, you see.

— Lewis Carroll, *The Hunting of the Snark: An Agony in Eight Fits*

Chapter 8

SUMMARY AND CONCLUSION

Ah, but a man's reach should exceed his grasp.

Or what's a heaven for?

— *Robert Browning*

This thesis has been concerned with the theory and application of VHF wind profilers to studies of the lower atmosphere. In particular, we have concentrated on the contribution which wind profilers can make to regional meteorology and operational forecasting, and the limitations which they experience in such applications. To this end, we have considered a number of examples of the profiler's main attributes: the ability to measure purely-vertical profiles of atmospheric structure and stability; and the potential to combine such profiles at a high temporal resolution in order to display accurate time-height cross-sections at a point in space. We have also concentrated on comparing profiler results with standard techniques, where appropriate. The profiler has major advantages over conventional instruments and methods in a number of ways, generally based on the excellent time resolution which the profiler allows. We have also highlighted the additional insights which wind profilers can deliver, both to applied forecasting and meteorology as mentioned above, and to general studies of atmospheric structure.

A number of investigations made over the course of these studies, related to more technical aspects of the profiler and associated atmospheric measuring systems, are

presented in Chapter 3. We have considered the effects of random errors in the profiler antenna phasing systems, and found that the effect on beam pointing direction (and hence the derived wind) can be significant. Generally, a well-maintained radar will have errors similar to those associated with factors like ground tilts (around a tenth of a degree). The effects on (horizontal) winds determined by well off-vertical Doppler beams is negligible, but estimates of vertical velocity (radial velocity on a vertical Doppler beam) can be affected by about 0.15% of the horizontal wind folding into the vertical-beam estimate. These random phasing errors can also affect the overall beam pattern, with increasing random phase and magnitude errors producing a decrease in sidelobe suppression, and a resultant increase in isotropic (all-sky) returns. Array phase measurements are not common, and our results indicate that profiler maintenance should include retuning of the array elements on a regular basis to ensure the integrity of the radar beam shape and pointing angle. This is especially important for interferometry measurements, which are becoming increasingly popular as a research technique. Variations in measured phase can only be assigned to an atmospheric cause after instrumental biases (including those due to environmental effects) are removed.

Extended comparisons between spaced antenna winds and colocated windfinding radiosonde retrievals have not been undertaken before, and profiler-sonde comparisons making use of large datasets are infrequent. We make an attempt at redressing this, as well as investigating experimentally the effects of inadequate sampling in the spaced antenna method. This problem has not been explicitly examined before, and we show how wind estimates can be affected in such a way as to be difficult to detect from a small number of measurements. The large number of individual comparisons we present makes the effects obvious, although there is still a wide range where it is hard to distinguish between accurate and inaccurate measurements. The spaced antenna wind underestimation reported in the literature at higher velocities may well be due to this effect, although further theoretical and experimental confirmation is required. Outside of the sampling-time affected region, the wind comparison is excellent. In particular, the application of a sliding time average to correct for temporal variations is shown to improve the comparison, albeit to a small degree. Further comparisons, especially

over a wide range of meteorological conditions such as we consider here, are essential to confirm the applicability of the spaced antenna technique to the measurement of lower-atmosphere winds. Extending the temporal correction we make to deal with *spatial* variations in the wind field (by considering horizontal wind gradients in detail, for example) would be particularly valuable.

Vertical velocity is one of the most important fields as far as weather forecasting is concerned, and the profiler's ability to make a direct measurement is highlighted throughout the thesis (during case studies in particular). It has previously been suggested that atmospheric vertical velocity may be extracted from the ascent rate of radiosondes, and we investigate this claim in some detail. We show that variations in ascent rate can be consistent with variations in drag coefficient throughout the ascent, and as such should be treated with some caution. The shape of the balloon can change as it ascends, producing feedback effects on the ascent rate; similarly, changes in the Reynolds number of the flow can be significant. Strong lee wave activity, with perturbations of the order of a couple of metres per second, is likely to overcome these variations, but it is far from certain that measurements in the "clear" atmosphere can be unambiguously assigned to atmospheric vertical velocity perturbations. The possibilities of feedback effects are intriguing, especially with regards to induced oscillations in the balloon shape under acceleration, and are deserving of further study. It may well prove that useful information, distinct from the vertical air velocity alone, can be extracted from the balloon ascent profiles in all conditions.

We move on to an examination of atmospheric vertical structure as observed by VHF profilers in Chapter 4, by making the first generally-applicable quantitative estimations of profiler height coverage; previous work has been exclusively qualitative, or confined to a particular locale. By making use of historical (or climatological) radiosonde thermodynamic data, we formulate a model and predictive algorithm for profiler signal-to-noise ratios as a function of altitude. This model is shown to be consistent with the observations at Adelaide (in the midlatitudes) and, after appropriate modifications based on observing system differences, independently applicable at Darwin (a tropical site). We show how the model predictions can be adjusted to take account of hardware and software concerns, so that profiler performance can be

predicted at any location for which long-term thermodynamic profiles are available. As an additional benefit, profiler range gates can be calibrated directly through such comparisons.

The long-term comparisons of profiler returned power with the generalised gradient of atmospheric refractive index, M , reveal variations in the range dependence of echo strength over altitude and time. Hence, it is possible to draw some conclusions about the varying nature of the atmospheric radiowave scatterers, by deriving a profile of the range-dependence exponent from the returned-power/ M^2 ratio. Recent moves by international weather organisations to archive high-resolution sonde data will be of major assistance to global studies of this variability, and related studies of atmospheric gravity wave activity. The parameterisation of gravity wave activity is of great interest to synoptic meteorologists, as breaking gravity waves exert a drag on the atmosphere which should be incorporated into global forecasting models. Our observed variations between morning and evening observations, and across seasons, are quite relevant here, implying that attention should be paid to temporal as well as spatial variations in gravity wave activity.

While local conditions such as the tropical monsoon play a part in predictions of profiler performance, generally it is the characteristics of the tropopause region which limit altitude coverage. In Chapter 5 we extend the ideas discussed above to real-time estimates of the tropopause height from the radar returned power profiles. Despite complications due to the same range-dependence problem considered earlier, the algorithm we implement demonstrates excellent performance over a wide range of conditions. There is some underestimation compared to sonde retrievals, which we associate with errors in assuming a range dependence going as r^2 across all altitudes. The algorithm can also misidentify enhancements due to moisture rather than (potential) temperature gradient, although this is obviously confined to the lower troposphere. We introduce a cluster-mean approach to the objective identification and tracking of multiple stable layers in the lower atmosphere, and show that the strong layers seen by the profiler have lifetimes in the order of hours to days; the lifetime can be used as a means of discriminating between tropopause-like layers and moisture enhancements. Some speculative possibilities for an extension of the technique to the

retrieval of temperature and humidity profiles are discussed, although the latter suffer severely from poor profiler range resolution and a difficulty in discriminating explicitly between moisture- and temperature-induced fluctuations.

The question of varying range-dependence is an important one, and has implications to all techniques which make use of scattering comparisons from the lower stratosphere in particular. We have seen this in both of the studies discussed above: calculations of M^2 overestimate the range-corrected (r^2) returned power in the lower stratosphere on many occasions, implying that a height-variable exponent greater than 2 is appropriate in that region; and estimation of the tropopause height is biased by a regular inability to accurately calculate the average pressure- and range-corrected returned power above the tropopause. Although there is a seasonal variation in aspect sensitivity which can contribute to a long-term mean range-dependence exponent, the detailed case studies we make show that local variations due to weather effects in the lower atmosphere (which affect the lower stratosphere through the propagation of gravity waves, for example) can dominate short-term results. An important continuance of the work we present here will be a study of aspect sensitivity variations in the lower stratosphere, especially with regards to the source of such variations. Although short-term weather-induced fluctuations may dominate, the inclusion of a climatological mean range-dependence exponent profile into experiments similar to those in Chapters 4 and 5 would be another step forwards in our understanding of the nature of scattering processes in the lower atmosphere.

Having examined vertical structure as probed by the wind profiler, we move on to mesoscale meteorological applications of the instrument and technique. In Chapter 6 we make a detailed investigation of an intense midlatitude cyclone, which produced heavy rain undiagnosed by local forecasting efforts. The additional insights into this particular system which we present include its identification as an interaction between the subtropical front and a Southern Ocean cold-air vortex. In particular, it appears to be a Southern Hemisphere example of the “instant occlusion” phenomenon which has previously been reported in the Northern Hemisphere. We have detailed the explicit pattern of air flows around the system, and how these flows combine with the available moisture to produce a destabilisation of the atmosphere as cold, dry

air overruns warm, moist air, resulting in the triggering of severe convection. By identifying this mechanism to forecasters, it is hoped that the accuracy of forecasts in similar situations can be improved: most forecasting errors are associated with a lack of previous case studies upon which to base deductions from a given situation.

The striated delta cloud depicted in Plate I suggests that a further avenue of research is in gravity wave generation by systems such as this. A process of geostrophic adjustment suggests itself, whereby deceleration of the mean flow in the diffluent region near a jet exit ahead of the system is associated with energy transfer and deposition by the radiation of gravity waves. An examination of radiosonde ascents in the vicinity of this cloud would likely prove illuminating, and is a projected extension of the case study.

The combination of RASP mesoscale analyses, radiosonde ascents, and wind profiler cross-sections allows us to make intercomparisons of the techniques. We have confirmed the self-consistency of the dataset with respect to these techniques, including a comparison of VHF vertical velocities with the RASP analyses. This is a particularly valuable contribution, since intercomparisons of vertical velocity estimates are few and far between in the literature. By confirming agreement between the qualitative pictures, and making significant steps towards a quantitative comparison, confidence in the accuracy and applicability of profiler vertical velocity measurements is enhanced. This is especially true since our comparison is made during the passage of a strong mesoscale system, with strong changes in the three-dimensional wind field throughout the case study. The time-height and longitude-height cross-sections agree well, especially when factors such as temporal evolution are taken into account. Future comparisons with higher-resolution mesoscale models will allow a more accurate quantitative comparison.

The second major case study we present concerns the behaviour of frontal systems during frontogenesis in the vicinity of a prefrontal trough. This is a relatively common occurrence in the southern Australian region, and is a continual source of forecasting error to local meteorologists. We show that the inclination for forecasters to identify cloud bands with active cold fronts is not always correct, especially when this inclination overrides conventional surface-observation based identification

of frontal zones. By showing that the arrival of the prefrontal trough coincides with the observed temperature and wind change at the surface, while the analysed “surface cold front” creates barely a ripple in the observations, we have confirmed (at least for this case) the model of frontal movement by propagation rather than advection. The profiler observations of vertical velocity provide insights into the location of cloud development which are similarly consistent with the model of a single change line at the position of the “prefrontal” trough. Further work on this important local feature is required, however, in an effort to tie down the motion of the frontogenetic region in time and space: high-resolution RASP analyses are an obvious starting point, combined with profiler and sonde measurements as we have already demonstrated. We recommend that the relative vorticity field should be a first avenue of investigation, as it can aid in identification of frontal zones and other regions of development (as seen in Chapter 6).

In conclusion, we have demonstrated the applicability of wind profilers to studies of lower atmosphere structure and mesoscale meteorology. We note that the Buckland Park profiler has been used regularly by the local Bureau of Meteorology Regional Office as an aid to forecasting, with products such as vertical velocity and radar tropopause heights made available through analysis and interface systems designed by the author of this thesis. The ability to “nowcast” via dialup access to the profiler has been used extensively by the forecasting staff to confirm predictions, and on occasion to assist in forecasts (W. Grace, personal communication 1994). As the staff become more familiar with the profiler measurements, it is expected that the use of such data in forecasting will increase. Profilers provide an excellent adjunct to standard radiosonde measurements, offering continual, real-time measurements of important atmospheric parameters such as vertical velocity. With the ability to study the more basic physics of the atmosphere on vertical scales of better than a few hundred metres, wind profilers are a valuable research tool in their own right; combined with other techniques, their unique capabilities serve to improve our knowledge of the atmosphere and atmospheric processes as a whole.

For I dipt into the future, far as human eye could see,
Saw the Vision of the World, and all the wonder that would be;
Saw the heavens fill with commerce, argosies of magic sails.
Pilots of the purple twilight, dropping down with costly bales;
Heard the heavens fill with shouting, and there rain'd a ghastly dew
From the nations' airy navies grappling in the central blue:
Far along the world-wide whisper of the south-wind rushing warm,
With the standards of the peoples plunging thro' the thunder-storm;
Till the war-drum throb'd no longer, and the battle flags were furl'd
In the Parliament of Man, the Federation of the World.
There the common sense of most shall hold a fretful realm in awe,
And the kindly earth shall slumber, lapt in universal law.

—Alfred Lord Tennyson, *Locksley Hall*

Appendix A

ANTENNA SWITCHING CONTROL

The antenna switching hardware was originally developed by *Lawry* [1986]; however, it has undergone at least two significant modifications since that time. This report is an effort to summarise the current state of operation of the “black box”.

In the standard arrangement, the radar hardware and software interacts with the transmission and reception gear in the caravan via the radar controller. The antenna switching hardware (hereinafter referred to as “the box”) is placed between the radar controller output and the lines to the caravan, and acts to modify the controller outputs. One can then set up a number of modes which would not normally be available.

A diagram of the circuit presently in the box is included with this report (see Figure A.1). The radar controller outputs two dual-line signals, $BS0\pm$ and $BS1\pm$; the caravan hardware is looking for one of three cases:

<i>Input to Caravan</i>	<i>Beam Direction</i>
(0,0)	Vertical
(0,1)	West
(1,0)	East

where the input is given in terms of the DS8820 output on controller lines ($BS0+$, $BS1+$). The basis of operation of the box is to force line $BS1+$ to the caravan low, and use line $BS0+$ as the receiver array switching code. Tracing through this operation:

1. Controller lines BS0 $\pm$ are passed directly to the caravan;
2. Caravan line BS1+ is forced low (hence forcing the inverted line BS1– high), by the DS8830 dual line driver. This is equivalent to a forced low on the output from a DS8820 dual line receiver at the caravan end;
3. The previous two points mean that only two beam directions are possible (vertical and east, from Table A);
4. The choice of receiver array is also made from the controller lines. Both pairs are passed through a DS8820, and the resulting two outputs are fed to a standard 74LS00N 2-input NAND gate;
5. The output state of the NAND is used to drive a transistor switch, which controls the Midtex Type 190 relay. With the current wiring, a controller output of (0,0) results in the relay connecting the YAGI to Rx0, whilst controller outputs of (0,1) or (1,0) connect the COCO to Rx0.

Bringing this all together, we can arrive at a truth table for the operation of the box. It is easiest to look at the result for each mode of operation: that is, if we direct the controller (via software) to transmit on some mode, what is the actual result when the box is connected?

<i>Mode Selected</i>	<i>Beam Direction</i>	<i>Reception Array</i>
Vertical	Vertical	YAGI
West	Vertical	COCO
East	East	COCO

Note that the reception array result can be reversed (YAGI $\leftrightarrow$ COCO) by simply reversing the wiring on the relay.

As a final comment, it would be useful if these functions could be implemented in software. At present, to either connect or disconnect the box requires physically changing cables. This could easily be modified to switch operation, which could then be run from the software. However, this would require another line from the controller dedicated to the state of the box connection (in use/not in use), and such a line is

currently not available. The proposed upgrade of the VHF Radar Data Acquisition System (RDAS) to four (or more) channels would be an alternate solution.

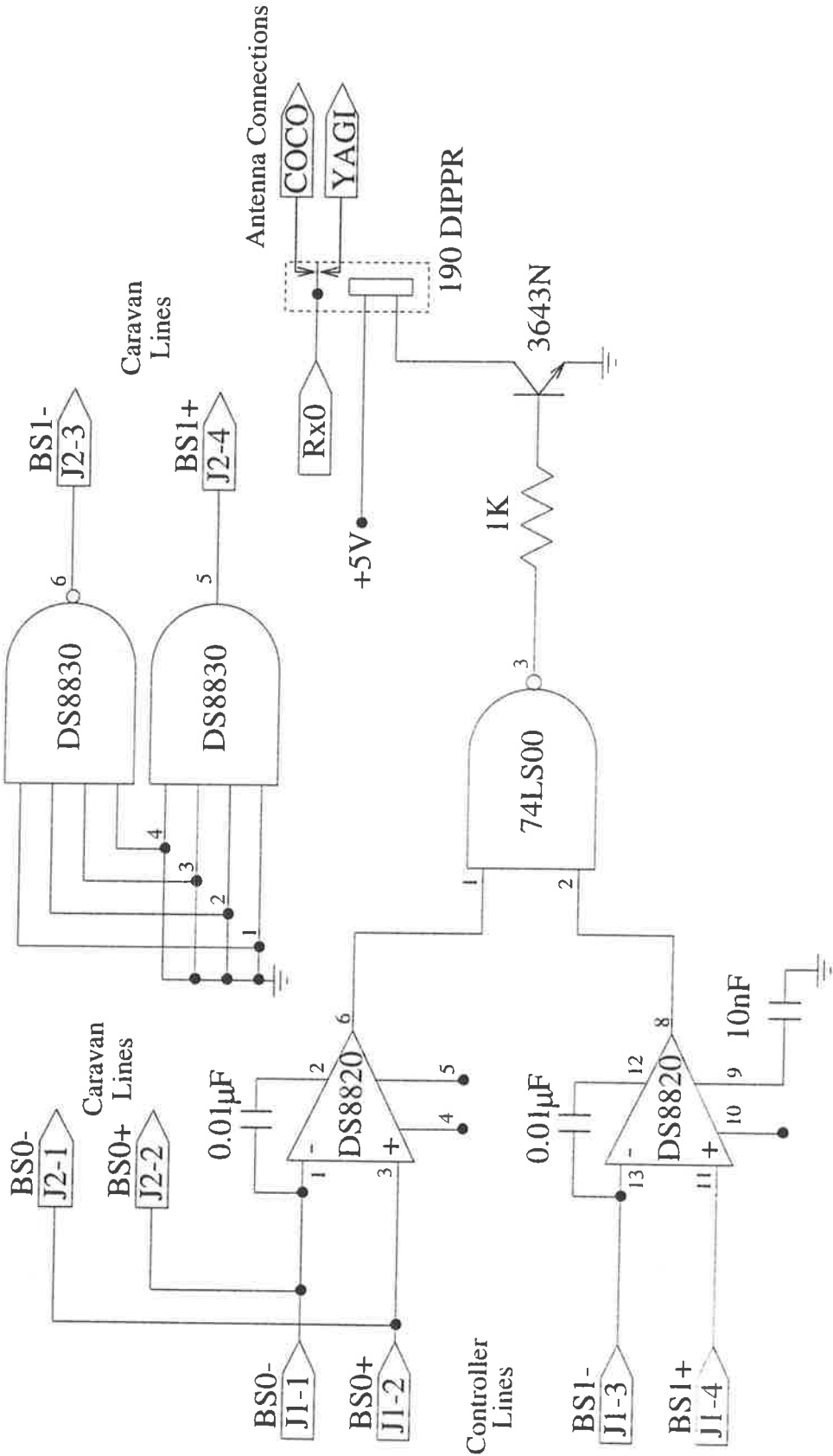


Figure A.1: Antenna switching circuitry.

Appendix B

COCO ARRAY POLAR DIAGRAM MODEL

B.1 Outline and Implementation

As described in Chapter 3, the Buckland Park VHF main array comprises 32 coaxial-collinear (CoCo) dipole rows, each of 48 elements. The rows are spaced $\lambda/2$ apart, with a “missing” central row (that is, if we number the rows from 0 to 31, there is a spacing of λ between rows 15 and 16). Figure 3.1 shows a schematic of the array. We wish to model the polar diagram produced by such an array, and examine the effects of various parameters.

Classical wave theory would see us combine individual wave components from each dipole element [Kraus, 1950; Balsley & Ecklund, 1972]; our approach is to model the array via Fourier theory. These methods are equivalent (see, for example, Bracewell [1986]), although there are some computational and conceptual advantages to the Fourier approach which are achieved at the expense of greater numerical precision. As an example, the Fourier approach we employ is about an order of magnitude faster than an equivalent summed component model; however, increasing the resolution of the Fourier approach (and hence the resolution of any cross-section through the resultant array polar diagram) rapidly erodes this advantage [Press *et al.*, 1992].

The first step is to model a row as a set of 48 dipole elements (radiators) spaced

at $\lambda/2$ along a line. Note that the cable space factor ($2/3$ in our case) will mean that the *physical* length of the antenna row is different to the *electrical* length, measuring in wavelength units. However, the physical aperture size is the important factor as far as antenna beam-forming is concerned: each row is 24λ (in cable) long, or 16λ (about 89 m) in free space.

Electrical losses along the cable amount to some 5.18373 dB per 100 metres. Since each row is centre-fed, this can easily be accounted for by adjusting the magnitude of each complex element along the row, away from the centre. Additionally, there is a triplet factor along a CoCo row due to self-interactions [Judas, 1987], such that each set of three elements progressing away from the centre-feed point are in the power ratio of (approximately) $[1 : 2 : 1]$; there are eight such triplets along each half of a CoCo row.

We can now move on to considering the differences between rows. Firstly, there are increasing losses from row to row, away from the centre of the CoCo array, due to increasing cable feed lengths. Feed cables are cut to measure an integral number of wavelengths to maintain phase consistency, but there is a difference of up to ten wavelengths in physical length between the rows closest to the centre and those on the outside of the array. These losses can be treated in precisely the same manner as the losses along each row, altering the “starting point” magnitude at each centre-feed point. It is worth noting at this point that cable losses together with a centre-feed approach serve to enhance apodisation effects, reducing the magnitude of sidelobes by limiting the amount of power at the edges of the array aperture. This occurs in both dimensions, along and across rows.

Since the Buckland Park profiler beam can be swung in the E-W plane, beam pointing was included in the model; this is simply a case of incrementing the phase of the complex signal at the feed-point of each row (the shift theorem of Fourier theory). Random phase errors can be similarly introduced at this point. Finally, random magnitude variations can be applied.

It is then simply a case of arranging the dipole rows into an array, and performing a two-dimensional Fourier transform. Our arrangement is depicted in Figure B.1. Note that the array is physically 16λ square, and that elements are separated by $\lambda/2$ in both

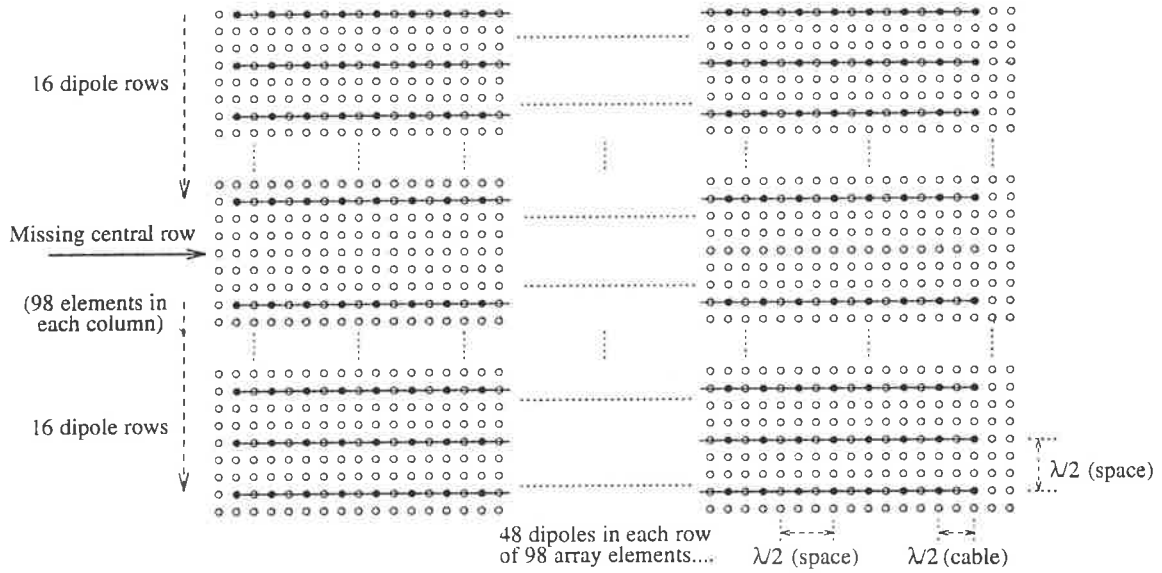


Figure B.1: A template of the modelled array aperture function. Filled circles denote the positions of dipoles, open circles are empty “filler” space in the numerical array; cables positions are indicated by solid lines. There are 48 dipole elements along each cable row, and 16 cable rows each side of the missing central row. The mathematical array is 98 elements square. Open circles are assigned a value of 0, and filled circles are initially given values of 1; each step in the model serves to modify the values at the positions of the filled circles.

directions; however, since the “along-row” dimension is within cable, the apparent physical spacing in that direction is less. We simulate this using the convenient value of the cable space factor ($2/3$), by arranging the dipoles every second unit along rows, and arranging the rows to be spaced by three units. Note that the central row is missing. The mathematical array is kept square, with an even number of elements (98) in each direction, to retain symmetry. We zero-pad this 98-element square array by centering it within a 1024-element array, a technique designed to reduce edge effects in the Fast Fourier Transform (FFT). Performing an FFT on the entire array—that is, applying a Fourier Transform to the aperture—gives as a result the distribution of power around the sky. Multiple realisations of the FFT can be made to simulate the effects of random phase errors, giving an average power spectrum.

B.2 IDL Code: fft-polar.pro

A copy of the IDL code software used in the array modelling is included here for reference.

```

pro choose,choice
    print,'Make a choice....?'
    print,'0: Set up plotting parameters'
    print,'1: Calculate polar diagram'
    print,'2: Plot polar diagram (cross-section)'
    print,'3: Plot polar diagram (contour)'
    print,'4: Calculate beam widths'
    read,choice
end

choose,choice
case choice of
    0: begin
        on_error,2
        resetplt,/all
        !p.charsize=1.5
        !p.region=[0.05,0.05,0.95,0.95]
        !x.margin=[2,2]
        !y.margin=[3,0]
        !p.title='!17VHF CoCo Array Polar Diagram'
        !x.style=1
        !x.range=[-15,15]
        !x.minor=5
        !x.ticks=6
        !x.title='Zenith Angle (degrees)'
        !y.style=1
        !y.range=[-25,0]
        !y.minor=5
        !y.ticks=5
        !y.title='Relative Power (dB)'
    end
    1: begin
        lossfac=-0.0518373                ;dB per metre of coax
        lambda=5.545                      ;free-space wavelength in metres
        space=2.0/3.0                     ;cable space factor
        arsize=1024
        arsize2=arsize/2
        nrows=33                          ;number of CoCo rows (32, +1 "missing" row)
        missing=16                         ;the missing row!
        elements=48                       ;number of dipole elements along each row
        along_row=(0.5+findgen(elements/2))/2.0*lambda*space    ;positions

```

```

row_loss=10^(along_row*lossfac/10.0)
row_loss=[reverse(row_loss),row_loss] ;line losses along rows
col_length=fix((0.25+(findgen(16)+1.0)/2.0)/space+1)*lambda*space
col_loss=10^(col_length*lossfac/10.0)
col_loss=[reverse(col_loss),1.0,col_loss] ;line losses across rows
triplet=[1,2,1,1,2,1,1,2,1,1,2,1,1,2,1,1]
triplet=[triplet,triplet]
triplet=[triplet,triplet]
triplet=[triplet,triplet]
triplet=float(triplet) ;CoCo dipole triplet effect
nel=98 ;number of elements in the model array
row=complexarr(nel)
row_elt=indgen(elements)*2+1 ;position of each element in the array
for i=0,(elements-1) do row(row_elt(i))=triplet(i)*row_loss(i) ;values along each row
col_elt=indgen(nrows)*3
totalpd=complexarr(arsize,arsize)
totalpd(*,*)=0.0
rndth=replicate(0.0,nrows)

print,'Zenith (deg,+W), phase S.D. (deg), mag S.D. (%), iterations'
read,cent,var,percent,numits
percent=percent/100.0

for j=1,numits do begin
  print,'Starting iteration number ',j
  array=complexarr(arsize,arsize)
  array(*,*)=0.0
  arrpd=replicate(0.0,arsize,arsize)
  rotarr=findgen(nrows)*cent*!pi ;beam tilt
  rndth=var*randomn(s,nrows) ;random errors in phases
  fullarr=rotarr+rndth
  rndmag=replicate(0.0,nrows)
  rndmag=percent*randomn(s,nrows)+1.0 ;random errors in magnitudes
  sinrnd=sin(fullarr*!dtr)
  cosrnd=cos(fullarr*!dtr)
  aperture=complexarr(nel,nel)
  aperture(*,*)=0.0
  for i=0,(nrows-1) do begin
    temp=reform(row*col_loss(i)*rndmag(i)*complex(cosrnd(i),sinrnd(i)))
    aperture(col_elt(i),*)=temp ;fill in all the rows
  endfor
  aperture(col_elt(missing),*)=0.0 ;centre row is missing

;Now, embed the aperture within a zero-field for the purposes of the FFT...
array((arsize2-nel/2):(arsize2+nel/2-1),(arsize2-nel/2):(arsize2+nel/2-1))=aperture

arrpd=abs(shift(fft(array,-1),arsize2,arsize2))^2 ;calculate the FFT

```

```

        totalpd=totalpd+arrpd                                ;sum realisations
    endfor

    x=findgen(arsize)/float(arsize2)-1.0                      ;convert units to degrees
    sinth=x
    rx=asin(x)
    costh=cos(rx)
    dx=(3.0)*rx/!dior
;Need to multiply by (3.0) since three "units" in the model equals a lambda/2 physical spacing

    avpd=10.0*log10(abs(totalpd)/float(numits)+1.0e-20)
    avpd=avpd-max(avpd)                                       ;rescale the average polar diagram to zero-max

    xsec=fix(arsize2*(sin(cent*!dior/3.0)+1)) ;where to take the NS cross section
end
2: begin
    plot,dx,avpd(xsec,*),linestyle=2,psym=-2 ;N-S
    oplot,dx,avpd(*,arsize2)                ;E-W
end
3: begin
    contour,avpd,dx,dx,yrange=[-10,10],xrange=[-10,10],xminor=5,xticks=4, $
        yminor=5, yticks=4, xtitle='Degrees E-W', ytitle='Degrees N-S', $
        levels=[-25.0,-20.0,-15.0,-10.0,-5.0,-3.0,-1.0], c_labels=[1,0,1,1,1,1,1]
end
4: begin
    guess=xsec
    repeat begin
        guess=guess-1
    endrep until ((avpd(guess,arsize2) lt -3.0) and (avpd(guess+1,arsize2) gt -3.0))
    m=avpd(guess+1,arsize2)-avpd(guess,arsize2)
    c=avpd(guess,arsize2)-m*guess
    ewwidth=(-3.0-c)/m                                     ;linear fit to find width
    ewwidth=3.0*asin((arsize2*(sin(cent*!dior/3.0)+1)-ewwidth)/float(arsize2))/!dior ;in degrees
;And repeat for the N-S cross-section...
    guess=arsize2
    repeat begin
        guess=guess-1
    endrep until ((avpd(xsec,guess) lt -3.0) and (avpd(xsec,guess+1) gt -3.0))
    m=avpd(xsec,guess+1)-avpd(xsec,guess)
    c=avpd(xsec,guess)-m*guess
    nswidth=(-3.0-c)/m
    nswidth=3.0*asin((float(arsize2)-nswidth)/float(arsize2))/!dior
    print,ewwidth,nswidth
end
endcase
end

```

Appendix C

THE SAFEx PROGRAMME

This appendix gives an overview of the South Australian Fronts Experiment (SAFEx), a collaborative observing programme carried out by the University of Adelaide's Department of Physics and Mathematical Physics, and the Centre for Dynamical Meteorology and Oceanography at Monash University. Väisälä MARWINsonde measurements of pressure, temperature, dewpoint, humidity and mixing ratio, together with zonal and meridional wind components, were combined with 54.1 MHz wind profiler observations of the reflectivity and three-dimensional wind field. Field measurements were taken at the Buckland Park station, some 40 km north of Adelaide. Other features of the experiment have already been discussed.

SAFEx was conducted in four Phases (I, II, III and IV) during 1991 and 1992; each Phase comprised one or more Intensive Observation Periods (IOPs) which combined high-resolution radar, radiosonde and AWS data. The aim was to observe a variety of frontal-type events with detail unprecedented in the region. Although nearly twenty events were catalogued, there were very few "strong" systems (involving large temperature and/or wind changes) which passed over Adelaide during the observing periods; many frontal systems tracked south of the continent. The strong El Nino phase throughout the SAFEx timeframe was almost certainly responsible for this perturbation. For example, *Karoly* [1989] has shown that a high pressure anomaly is present over Australia during the El Nino, which is consistent with the southward displacement of mesoscale systems.

A brief survey of the entire SAFEx programme is undertaken below. Times are quoted below as UTC; Adelaide Local Standard Time (LST) is (UTC+9.5 hours), except for the “daylight savings” period through October–March when it is (UTC+10.5 hours). The sheer quantity of high-quality data obtained through SAFEx ensures that it cannot all be analysed in a single work such as this. Indeed, the preliminary working analyses generated through the tenure of this thesis, comprising fields similar to those presented earlier but for each and every recorded event and “inter-event” period, fill more than a metre of shelf-space in the author’s bookcase. We have endeavoured to highlight some of the most interesting cases (Chapters 6 and 7) together with some general overviews, but there remain significant opportunities for extended studies with this comprehensive dataset.

C.1 Phase I: June 1991

C.1.1 IOP I/1

This event consisted of a weak surface front, due to movement of the main system towards the south. A strong upper trough and prominent North-west cloudband are notable features. The wind profiler (8kW peak output power) alternated vertical Doppler (2 km–16.5 km) with spaced antenna (2 km–11.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.1: Release times (UTC) of the fifteen (15) balloons launched during IOP I/1.

UTC Date	Launch Times
19 June 1991:	0829 1130 1431 1731 2030 2330
20 June 1991:	0230 0530 0830 1130 1435 1733 2030 2330
21 June 1991:	0230

C.1.2 Non-IOP flights in Phase I

Table C.2: Release times (UTC) of the non-IOP balloons launched during Phase I.

UTC Date	Launch Times
17 June 1991	2304
18 June 1991	2301
26 June 1991	0509

C.2 Phase II: December 1991

C.2.1 IOP II/1

Although the surface front was shallow, the upper front in this IOP was significant. The surface front was associated with a good temperature and wind change, and a prominent cold dome in the lower troposphere. The wind profiler alternated vertical Doppler (2 km–16.5 km) with spaced antenna (2 km–11.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.3: Release times (UTC) of the seven (7) balloons launched during IOP II/1.

UTC Date	Launch Times
2 December 1991:	0006 0203 0414 0558 0801 1008 1206

C.2.2 IOP II/2

A rather weak event. The wind profiler alternated vertical Doppler (2 km–16.5 km) with spaced antenna (2 km–11.5 km) on a five-minute cycle at 500 m sampling throughout the IOP, although observations were somewhat patchy.

Table C.4: Release times (UTC) of the three (3) balloons launched during IOP II/2.

UTC Date	Launch Times
4 December 1991:	2159
5 December 1991:	0406 0955

C.2.3 Non-IOP flight in Phase II

Table C.5: Release time (UTC) of the non-IOP balloon launched during Phase II.

UTC Date	Launch Time
7 December 1991	0158

C.3 Phase III: February–March 1992

C.3.1 IOP III/1

A very strong surface front and associated prefrontal trough, with a temperature drop of 15°C over an hour. The wind profiler alternated vertical Doppler (2 km–16.5 km) with spaced antenna (2 km–11.5 km) on a five-minute cycle at 500 m sampling, but data was confined to the first six hours of observations due to technical difficulties with new power supplies (16 kW peak output power).

Table C.6: Release times (UTC) of the ten (10) balloons launched during IOP III/1.

UTC Date	Launch Times
16 February 1992	2302
17 February 1992	0105 0316 0507 0647 0837 1029 1247 1451 1722

C.3.2 IOP III/2

A typical summertime front, with a weak prefrontal trough. The wind profiler (8 kW peak output power) alternated vertical Doppler (2 km–16.5 km) with spaced antenna (2 km–11.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.7: Release times (UTC) of the ten (10) balloons launched during IOP III/2.

UTC Date	Launch Times
20 February 1992	2205
21 February 1992	0101 0359 0700 0951 1300 1552 1821 2136
22 February 1992	0128

C.3.3 IOP III/3

Another weak event; although the prefrontal trough appeared interesting, the event was undersampled. The wind profiler alternated vertical Doppler (2 km–16.5 km) with spaced antenna (2 km–11.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.8: Release times (UTC) of the three (3) balloons launched during IOP III/3.

UTC Date	Launch Times
3 March 1992	2334
4 March 1992	1440 1736

C.4 Phase IV: August–October 1992

C.4.1 IOP IV/1

Weak low-level front, but strong upper-level front. The wind profiler alternated vertical Doppler (2 km–16.5 km) with spaced antenna (2 km–11.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.9: Release times (UTC) of the ten (10) balloons launched during IOP IV/1.

UTC Date	Launch Times
18 August 1992	0659 1115
19 August 1992	1458 1901 2229 2356 ^a 0204 0440 0848 1432

^aNo winds for this flight

C.4.2 IOP IV/2

A promising event, with a good cold dome, which was undersampled. The wind profiler range was increased due to improved power output (approximately 16 kW peak available), and alternated vertical Doppler (2 km–19.5 km) with spaced antenna (2 km–14.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.10: Release times (UTC) of the four (4) balloons launched during IOP IV/2.

UTC Date	Launch Times
26 August 1992	1316 1920
27 August 1992	0529 1332

C.4.3 IOP IV/3 (and IOP IV/4)

The “Cudlee Creek storm”, as described in Chapter 6. Very heavy, intense rainfall associated with a cut-off low to the south of Adelaide was a major feature. The original IOP IV/3 (sixteen flights) was later combined with the then IOP IV/4 (six flights) to form what is now referred to as IOP IV/3. The wind profiler alternated vertical Doppler (2 km–19.5 km) with spaced antenna (2 km–14.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.11: Release times (UTC) of the eighteen (18) balloons launched during IOP IV/3.

UTC Date	Launch Times
28 August 1992	1449 1829
29 August 1992	0050 0203 0604 1047 1257 1628 2141
30 August 1992	0048 0325 0626
30 August 1992	1344 1714 2125
31 August 1992	0121 0434 0708 ^a

^awinds below 10 km only

C.4.4 IOP IV/5

A very weak, low-level change. The wind profiler alternated between vertical, west and east Doppler beams (2 km–19.5 km) on a six-minute cycle at 500 m sampling throughout IOP.

Table C.12: Release times (UTC) of the eleven (11) balloons launched during IOP IV/5.

UTC Date	Launch Times
5 September 1992	2046
6 September 1992	0013 0428 0740 1029 1331 1632 1924 2227
7 September 1992	0141 0557

C.4.5 IOP IV/6

A weak event with a wind change but no temperature signature. The wind profiler alternated between vertical, west and east Doppler beams (2 km–19.5 km) on a six-minute cycle at 500 m sampling until 0522 UTC 11 September, then vertical Doppler beam only (2 km–19.5 km) on a two-minute cycle at 500 m sampling for the remainder of the IOP.

Table C.13: Release times (UTC) of the eight (8) balloons launched during IOP IV/6.

UTC Date	Launch Times
10 September 1992	0247 0829 1228 1634 2038
11 September 1992	0129 0422 ^a 0844

^ano winds this flight

C.4.6 IOP IV/7

A strong upper front, with a weaker surface front. An excellent example of the pre-frontal trough interacting with a Southern Ocean cold front. The wind profiler collected vertical Doppler beam (2 km–16.5 km) data only, on a two-minute cycle at 500 m sampling throughout the IOP.

Table C.14: Release times (UTC) of the seventeen (17) balloons launched during IOP IV/7.

UTC Date	Launch Times
13 September 1992	0223 0834 1223 1623 1952 2252
14 September 1992	0129 0337 0526 0721 0925 1301 1627 2026
15 September 1992	0042 0430 0824

C.4.7 IOP IV/8

A weak front to the south of the continent associated with a prefrontal trough. The wind profiler alternated vertical Doppler (2 km–19.5 km) with spaced antenna (2 km–14.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.15: Release times (UTC) of the seven (7) balloons launched during IOP IV/8.

UTC Date	Launch Times
20 September 1992	1428 2035
21 September 1992	0230 0827 1446 2029
22 September 1992	0301

C.4.8 IOP IV/9

Another event with a good wind change, but little temperature change. A weak prefrontal trough was present. The wind profiler collected vertical Doppler beam data only from 0120 UTC to 0246 UTC 27 September (2 km–19.5 km, two-minute cycle, 500 m sampling), then alternated vertical Doppler (2 km–19.5 km) with spaced antenna (2 km–14.5 km) on a five-minute cycle at 500 m sampling for the remainder of the IOP.

Table C.16: Release times (UTC) of the eleven (11) balloons launched during IOP IV/9.

UTC Date	Launch Times
27 September 1992	0220 0658 1032 1329 1629 1940 2231 ^a
28 September 1992	0210 0530 0928 1322

^asporadic winds

C.4.9 IOP IV/S1

This was a comparison of nearly-simultaneous (interleaved) radiosonde, spaced antenna and Doppler beam-swinging winds. The wind profiler alternated spaced antenna (2 km–11.5 km) with vertical, east and west beam Doppler (2 km–16.5 km) on a nine-minute cycle throughout the IOP. Manual cable changes were made to overcome the three receiver channel problem.

Table C.17: Release times (UTC) of the two (2) balloons launched during IOP IV/S1.

UTC Date	Launch Times
1 October 1992	0137 0329

C.4.10 IOP IV/10

A weak front. The wind profiler alternated vertical Doppler (2 km–19.5 km) with spaced antenna (2 km–14.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.18: Release times (UTC) of the two (2) balloons launched during IOP IV/10.

UTC Date	Launch Times
5 October 1992	1219 1827

C.4.11 IOP IV/11

A strong event, but inclement conditions prevented more balloon launches. In this case, frontogenesis was actually analysed in a “prefrontal” trough from about halfway across the Bight. The wind profiler alternated vertical Doppler (2 km–19.5 km) with spaced antenna (2 km–14.5 km) on a five-minute cycle at 500 m sampling throughout the IOP.

Table C.19: Release times (UTC) of the four (4) balloons launched during IOP IV/11.

UTC Date	Launch Times
8 October 1992	0248 0851 1514 ^a 2105

^asporadic winds

C.4.12 Non-IOP flights in Phase IV

Table C.20: Release time (UTC) of the non-IOP balloon launched during Phase IV.

UTC Date	Launch Times
4 September 1992	0405 ^a

^avariously referred to as IOP IV/5a

Appendix D

SUMMARY OF SYMBOLS

A_e	=	antenna of effective area
a	=	universal constant ($\approx \sqrt{2.8}$)
B	=	spectral bandwidth
C	=	ground clutter
C_D	=	drag coefficient
C_n^2	=	turbulence refractive index structure constant
c	=	speed of light
c_{pd}	=	specific heat of dry air at constant pressure ($1005.7 \pm 2.5 \text{ J kg}^{-1} \text{ K}^{-1}$)
d	=	duty cycle ($= \Delta t/t_p = f_p \Delta t$)
$E_\zeta(k)$	=	spectrum of atmospheric Bragg-scale vertical displacements
e	=	water vapour pressure
$F_B(z)$	=	buoyancy force
F_D	=	drag force
F_W	=	weight force
$f(r)$	=	range dependence of the backscattered echoes
f	=	radar frequency
f_c	=	Coriolis parameter
f_D	=	Doppler shift
f_p	=	pulse repetition frequency, PRF ($= 1/t_p$)
G	=	antenna gain ($= 4\pi A_e/\lambda^2$)

$\mathcal{G}$	=	shear modulus ($= NkT \approx 0.3 \times 10^6$ Pa)
$g(z)$	=	acceleration due to gravity
H	=	(temperature-dependent) scale height of the atmosphere
k	=	vertical wavenumber
k_B	=	Boltzmann constant
l	=	antenna row separation ($= \lambda/2$)
l_t	=	correlation distance
M	=	vertical gradient of refractive index
M_d	=	“dry” part of M
M_n	=	n -th classical spectral moment
$\overline{m}$	=	molecular weight of air (28.8×10^{-3} kg)
N	=	Brunt-Väisälä (buoyancy) frequency
$\langle N \rangle$	=	noise power
$\overline{N}$	=	average background signal level
N_e	=	number density of free electrons
n	=	radio refractive index of the atmosphere
n_c	=	number of sequential samples
n_s	=	number of points
$\langle P \rangle$	=	signal power
$\mathcal{P}(r)$	=	profiler returned power
p	=	pressure
p_s	=	superpressure
$\mathcal{Q}(r)$	=	pressure-normalised returned power
q	=	water vapour mixing ratio (g kg^{-1})
R	=	gas constant
$\tilde{R}$	=	gas constant per mole ($8.3143 \text{ J K}^{-1} \text{ mol}^{-1}$)
$R(m)$	=	autocorrelation function
$R(z)$	=	balloon radius
R_0	=	initial balloon radius
R_d	=	gas constant in dry air ($287.04 \text{ J kg}^{-1} \text{ K}^{-1}$)

R_e	=	Reynolds number
R_u	=	unstretched ("just filled") balloon radius (≈ 0.285 m)
$\mathcal{R}(r)$	=	range-normalised received power
r	=	range
Δr	=	range resolution
r_F	=	Fresnel zone radius ($= \sqrt{r\lambda/2}$)
r_{trop}	=	tropopause altitude
$\mathcal{S}(r)$	=	signal to noise ratio
T	=	temperature
T_c	=	cosmic noise temperature
T_D	=	(absolute) dew point temperature
T_L	=	absolute temperature at the lifting condensation level
T_{rx}	=	receiver noise temperature
T_v	=	virtual temperature
Δt	=	pulse width
t_p	=	interpulse period, IPP
U	=	percentage relative humidity
$\mathbf{u}$	=	three-dimensional mean wind, (u, v, w)
$\mathbf{u}_h$	=	horizontal wind, (u, v)
u	=	zonal velocity
V_r	=	radial velocity ($= dr/dt$)
$V(z)$	=	balloon volume
v	=	meridional velocity
v_T	=	terminal velocity
W_f	=	width of the peak
w	=	vertical velocity
z	=	altitude
z_i	=	complex time series
α	=	radar antenna/transmission-line system efficiency (loss) factor
α'	=	a ratio of eddy diffusivities, usually taken to be unity

β	=	constant incorporating all of the radar and reflection constants
χ	=	zenith angle
δ	=	phase delay between adjacent CoCo antenna rows
ε	=	balloon material thickness ($\approx 0.15 \times 10^{-3}$ m)
η	=	volume reflectivity
η_v	=	viscosity
Γ	=	(environmental) lapse rate
Γ_d	=	dry adiabatic lapse rate
γ	=	constant ($= pV/T$)
κ_d	=	R_d/c_{pd} (0.2854 ± 0.0007)
λ	=	radar wavelength
∇	=	gradient operator, $(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial z})$
ω	=	vertical velocity in pressure coordinates ($= dp/dt$)
Φ	=	geopotential height
$\Phi_n(k)$	=	vertical wavenumber spectrum of generalised potential refractive index
$\phi(\tau)$	=	autocorrelation phase
ψ	=	azimuth angle
ρ	=	density
ϱ	=	amplitude reflection coefficient
$\rho(\tau)$	=	autocorrelation amplitude
σ	=	standard deviation
σ_c	=	static stability parameter constant
ς	=	scattering cross-section
τ	=	autocorrelation lag
θ	=	potential temperature
θ_e	=	equivalent potential temperature

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