THE NATURE OF PRECIPITATING LAYER CLOUDS AROUND THE WORLD

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Summary: The nature of precipitating layer clouds around the world is not well documented. In order to better quantify this, a synthesis of observations of these clouds from field experiments conducted around the world has been developed. Experiments considered have been conducted in Australia, Canada, China, Israel, Japan, Russia, Ukraine, United States, and several European countries. Extra-tropical cyclones were typically responsible for the production of the clouds, although in some cases these were modified by orography or were linked with mesoscale systems. Several features of the precipitating layer cloud systems are considered. These include cloud base and cloud top heights and temperatures, horizontal scales of cloud and precipitation, embedded phenomena such as rainbands, conveyor belt characteristics, and ice crystal and water droplet concentrations, and raindrop and ice crystal size distributions.

The results of the study illustrate quantitatively that the synoptic and mesoscale characteristics of precipitating layer clouds around the world are not strongly dependent on the geographic location, although the precipitation efficiency of the rainbands may be modulated by the nature of the subcloud layer and by orographic forcing.

The microphysical structure of the precipitating clouds is dominated by the ice phase in the clouds. The life time of liquid water above the melting layer is small away from regions of enhanced convection and orographic uplift. However, under some circumstances liquid water may presist near cloud top. There is conflicting evidence as to whether or not the number of ice crystals in layer clouds can be parameterised as a function of cloud top temperature. Observations suggest that for ice particles with sizes between 10 μm and 100 μm the distribution is best described by a power law and for sizes above 200μm the ice crystal size distribution is best described by a Marshall Palmer distribution.

Finally, the paper shows that there are extensive regions of the world where observations of precipitating clouds are lacking. This particularly applies to regions over the open sea.

1. INTRODUCTION

Recently, the Global Energy and Water Cycle Experiment (GEWEX) Cloud Science Team (*Browning*, 1994) identified one of the biggest challenges in the World Climate Research Programme as an improvement in the understanding of how the wide range of processes within clouds affects the atmosphere. The GEWEX Cloud System Study (GCSS) was set up to promote the description and understanding of key cloud system processes, with the aim of developing and improving the representation of cloud processes in general circulation models, GCMs. One of the types of cloud systems identified by the GCSS was precipitating layer cloud systems.

Precipitating layer cloud systems are a very commonly observed feature of the earth's climate. In the mid-latitudes these cloud systems are associated with frontal systems, depressions in the westerlies, lows cut-off from the westerlies and tropical extra-tropical cloud bands. They are strongly modulated by orography and may be associated with organised convection in the form of rainbands.

The precipitating layer clouds form an integral part of the global water and energy cycle that drives the earth's climate system. *Warren et al.* (1986) show that over the land the zonal mean of the average amount of altostratus cloud between 30° and 60° latitude varies from 15% to 25% in the Northern Hemisphere and from 16% to nearly 60% in the Southern Hemisphere. Over the oceans between 30° and 60° latitude the zonal mean altostratus cloud cover varies from 20% to 35% in both hemispheres and varies as a function of the season (*Warren et al.* 1988). The precipitating layer clouds are far from uniform in structure. For example, recently *Stephens and Greenwald* (1991) showed that the albedo and liquid water path is significantly different for mid-latitude oceanic clouds compared to clouds over the tropical oceans. *Heymsfield* (1993) in a review of the structure of clouds suggests that the water phases in a cloud will depend upon the temperature, the age of the cloud, the ambient aerosol population, composition and size as well as a host of other possible factors.

An inherent feature of the physical processes in cloud systems is the interaction between scales, which extend from the synoptic scale down to the microscale. Systematic field observations are needed across these scales in order to develop and validate cloud scale parameterisations. In the case of the parameterisation of precipitating layer clouds, validation is required both over the land and the sea. Where layer clouds are forming over the land, parameterisations need to be validated in the absence and presence of significant orography.

The first aim of this paper is to synthesise a conceptual picture of precipitating layer clouds based on observations of the common features and differences of precipitating clouds as found in various regions of the world. The second aim is to highlight those regions of the globe where observations are lacking. The third and final aim of the paper is to identify those observations that can be used to verify the cloud properties generated by general circulation models, (GCMs), and limited area models, (LAMs), and those observation that can be used to generalise the parameterisations of the microphysics used in GCMs and LAMs simulations.

This paper should be read in conjunction with the companion paper by *Stewart* (1994) on the critical processes within, and attributes of, frontal cloud. Two topics discussed by Stewart and not addressed in this paper are the importance of understanding the dynamics, and the moisture budgets of middle level layer cloud systems.

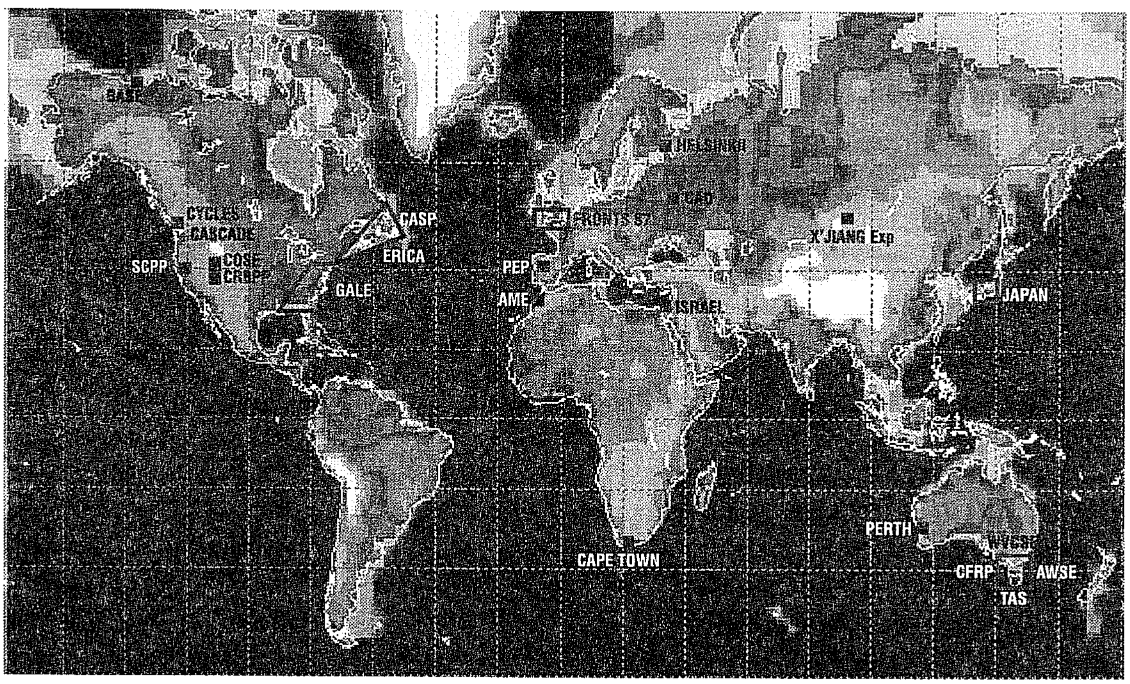

Fig. 1 A map showing the location of the major field experiments (Appendix 1), the major cloud seeding studies (Appendix 2) and the case studies discussed in the paper.

2. INTERNATIONAL STUDIES OF PRECIPITATING CLOUDS

Field studies of precipitating layer clouds over the past 20 years can be classified into three categories, namely (a) major meteorological observing programs, (b) major cloud seeding programs and (c) specific case studies. Major meteorological observing programs are co-ordinated studies of mid-latitude precipitating cloud systems. Examples of these types of studies are given in Appendix 1. Major cloud seeding studies, undertaken as part of a cloud seeding program usually focus on the microphysical properties of the clouds. Examples of major cloud seeding research studies are given in Appendix 2. The location of the major meteorological observing programs and the major cloud seeding programs that are relevant to the study of precipitating layer clouds are shown in Figure 1. Also shown in Figure 1 are some locations of case studies that are used in the paper. The case studies have been confined to regions where there are no reported major meteorological observing programs or major cloud seeding programs.

It is clear from Figure 1 that there have been extensive observational and cloud seeding programs that extend across of north America and western Europe. Over the land and in coastal regions there have been many case studies made in the United States, Canada, United Kingdom, France and Germany.

In southern Europe and the Middle-East cloud seeding studies have been undertaken in Spain, Morocco and Israel. In eastern Europe major field studies have been reported from Russia and the Ukraine (part of the old European U.S.S.R.) and in China a major cloud seeding study has been made in Xinjiang Province. In Japan several case studies provide observations for comparison with the European and north American studies. There are no significant observations over either the Atlantic or Pacific Oceans, although the data void over the Atlantic will be addressed during the Fronts and Atlantic Storm Tracks EXperiment, FASTEX (*Browning*, 1994).

In the Southern Hemisphere, the structure of precipitating layer clouds has been investigated as part of the Australian Cold Fronts Research Program (CFRP) and through studies made during cloud physics and seeding projects These studies focused on the southeastern corner of Australia, although there is one unpublished study made near Perth in Western Australia. Only one case study of precipitating layer clouds has been reported from South Africa. The Southern Alps Experiment (SALPEX) in New Zealand is a new initiative to investigate the clouds and precipitation upstream of and over the southern Alps. Apart from satellite observations there are no systematic studies over the vast ocean areas and papers on the structure of precipitating clouds in South America are lacking.

3 SYNOPTIC STRUCTURE OF PRECIPITATING CLOUDS

The precipitating layer clouds associated with fronts and mid-latitude cyclones are organised on the synoptic scale. If visualized along relative moist or dry isentropic surfaces, cloud formation (dissipation) is associated with regions of ascent (descent) of air within the sloping portions of the isentropic surfaces (see e.g. *Browning*, 1990 or *Carlson*, 1991). *Harrold* (1973) introduced the concept of the "conveyor belt" and identified the warm conveyor belt as the warm moist air stream which ascends relative to the warm front. *Carlson* (1980) extended these ideas to a conceptual model of the extra- tropical cyclone. The Carlson conceptual model consists of three streams of air. In the Northern Hemisphere, a warm conveyor belt enters the cyclone from the south or south-east rises in the warm sector and over the surface warm front, and exits the system to the north-east ahead of the upper-level trough. A cold conveyor belt approaches the cyclone in low levels from the north east, rises near the centre of the low and merges with the warm conveyor belt as it leaves the system towards the north- east. The third stream, the dry stream, originates in the upper troposphere upstream of the cyclone and descends into the system from the northwest (Figure 2).

The warm conveyor belt is the primary mid-level cloud producing flow. The conceptual model of the warm conveyor belt has been applied to account for aspects of cloud and precipitation distribution in north-western Europe (Harrold, 1973, *Browning* and *Monk*, 1982), the U.S. (*Carlson*, 1980) and Canada (*Stewart et al.*, 1988). In Australian studies of the summertime cool change (*Ryan* and *Wilson*, 1985) and wintertime cold fronts (*May et al.*, 1991) all three air flows were identified. Finally, *Matkovskii* and *Shakina* (1982) have identified all three airflows in an occluded front over the centre of the old European U.S.S.R.

In all of these studies the western and poleward edges of the cloud systems are well defined and the cloud bands follow the storm-relative isentropic streamlines. The cloud cover in the warm conveyor belt consists of mid- and high-level layers of altocumulus and altostratus or cirrus. The intrusion of dry upper tropospheric air influences the cloud pattern by generating moist instabilities in the warm conveyor belt. If the dry air undercuts the warm frontal zone or warm sector frontal zone it leads to evaporation of precipitation and instability in the subcloud layer (*Ryan et al.*, 1990) and if it over runs the warm conveyor belt it leads to the generation of a shallow moist zone with potential instability at cloud top (*Browning*, 1990).

The cloud structure below the warm conveyor belt is strongly dependent on the thermodynamic properties of the air and hence the geographic location. If the low-level air is subsiding (see *Browning*, 1990) or has risen from a dry continental region, such as occurs in southeastern Australia (*Ryan et al.*, 1989), precipitation from the overlying warm conveyor belt evaporates into the subsaturated cold conveyor belt. *Clough* and *Franks* (1991) point out that the ice and snow precipitation typically has a fall speed of the order of 1 m s^{-1} so may take about one hour to reach the ground. When the freezing level is well below cloud base substantial evaporation takes place before the snow begins to melt. The evaporating snow is

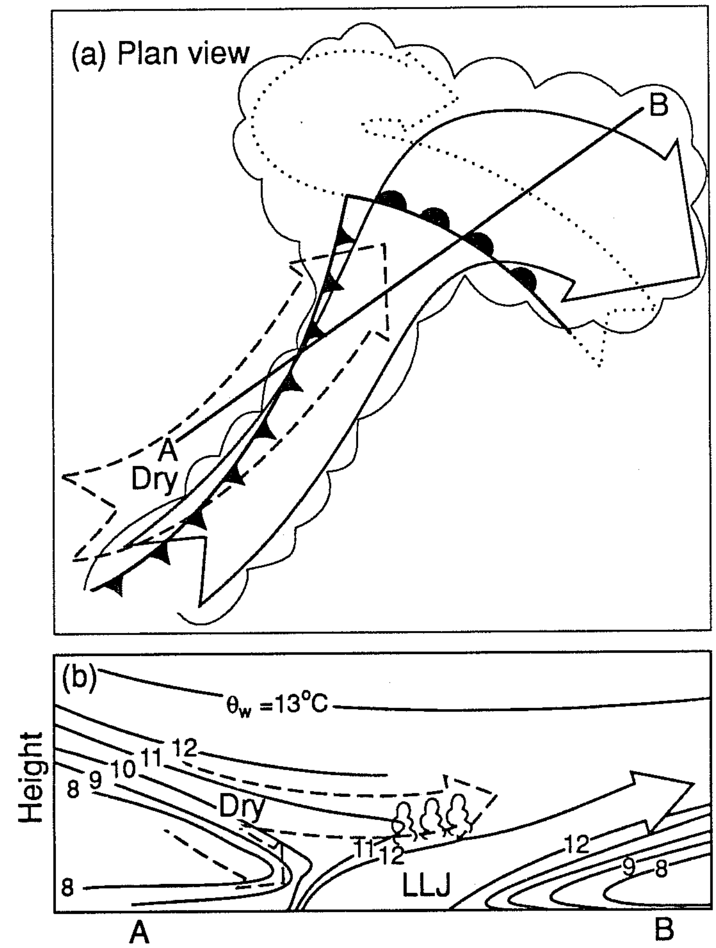

Fig. 2 A schematic portrayal of the airflow in an extra tropical cyclone in which the warm conveyor belt (solid arrow) is undergoing forward sloping ascent ahead of a kata- cold front, before rising above a flow of cold air ahead of the warm front. The dotted arrow is the cold conveyor belt and middle-tropospheric air is shown by a dashed arrow over running the cold front generating potential instability in the upper portions of the warm conveyor belt. (a) is a plan view and (b) is vertical section along line AB in (a). Flows are shown relative to the moving frontal system (From *Browning*, 1990).

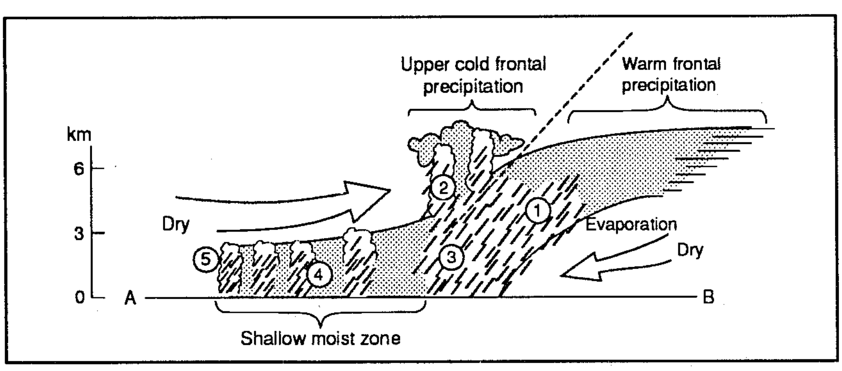

Fig. 3 A schematic representation of the regions of cloud and rain along section A-B in Figure 2. The numbers represent precipitation types as follows: (1) warm frontal precipitation; (2) convective precipitation and generating cells associated with an upper cold front; (3) precipitation from the upper cold front falling through the "feeder zone"; (4) shallow moist zone between the upper and surface cold fronts; and (5) shallow precipitation at the surface cold front (from *Browning*, 1990).

sufficient to maintain an air parcel descending at a rate of 10-30 cm s^{-1} at near-saturation whereas evaporating rain only maintains the parcel at about 40-60% saturation. Where the air beneath the warm conveyor belt rises and reaches saturation a deep layer cloud systems form. Under these conditions cloud base may be below one kilometer and there is little or no subcloud evaporation.

With the Russian observations made by *Matkovskii* and *Shakina* (1982), the warm conveyor belt has a long land trajectory with air parcels in the warm conveyor belt originating on the north coast of Africa, crossing the Mediterranean Sea and Turkey before entering eastern Europe. The cold conveyor belt enters the system from the east while the upper level belt of dry descending air comes from the north-west. As with the other regions of the world, Matkovskii and Shakina recognise that the warm conveyor belt and the descending dry air develop regularly irrespective of geographic region and season. However, the moisture content of the system, the presence of moist unstable layers and the degree of convection varies from case to case and is dependent on geography and season.

The warm conveyor belt and the accompanying precipitating layer cloud is typically several 1000 km long, a few 100 km wide and 1-2 km in depth (see e.g. *Harrold*, 1973). The scales of the conveyor belt flows are such that they are able to be resolved by limited area numerical models and general circulation models run at high resolution. Models with explicit liquid water schemes should be able to resolve the non-convective component of the conveyor belt cloud. However, as will be shown in the next section much of the cloud in the conveyor belt is coupled to mesoscale convective activity which is not explicitly resolved in either GCMs or LAMs.

4. MESOSCALE STRUCTURE OF PRECIPITATING CLOUDS

The vast majority of the recent observational studies of the mesoscale (20 to 200 km) structure of precipitating clouds have been associated with large field experiments requiring the synthesis of data from doppler radar, aircraft, radio sondes, drop sondes, wind profilers and satellite imagery. The majority these projects have taken place in north America, and western and eastern Europe. Most of the experiments identify the presence of rainbands or embedded convection. In this paper the term embedded convection is used when there is insufficient data to classify the characteristics of the rainband.

4.1 **Rainband Structure**

The current classification of rainbands is based largely on observational studies made in the United Kingdom and the west coast of the United States. These studies showed that frontal precipitation has an embedded banded structure (see e.g. *Houze* and *Hobbs*, 1982, *Browning*, 1985). The bands have been classified as narrow or wide rain bands according to their scale. The CYCLES observations and the U.K.

observations are both representative of rain bands associated with fronts that have formed between 40N and 50N in air of recent maritime origin.

Wide rain bands are typically 50-75 km across and are found embedded within and parallel to the frontal zone. *Matejka et al.* (1980) classified the wide bands into warm frontal bands, warm-sector bands, wide cold-frontal bands and pre-frontal cold surge bands. Narrow cold-frontal rainbands are 5-25km wide and tend to be generated by forced convection. They tend to occur in the cold season when the air ahead of the surface cold front has near neutral stability (*Browning*, 1990). Post-frontal rain bands form in the cold air behind the front.

An important characteristic of the warm frontal and warm sector clouds is the so-called snow "generating" cells first identified by "The McGill Stormy Weather Group" using RHI radar patterns (see e.g. *Douglas et al.*, 1957). Later *Wexler* and *Atlas* (1959) used combined radar and aircraft observations to show that the echo structure associated with the generating cells are embedded in the stratiform cloud deck. Subsequently, University of Washington group showed that the intense precipitation associated with wide rainbands appears to be associated with generating cells aloft. These generating cells seed the feeder layer clouds below (*Hobbs*, 1978).

In contrast to the wide rainbands, narrow cold-frontal rainbands are associated with a narrow updraft region, typically 5-25km in width. The updraft is coupled with a downdraft which is associated with heavy precipitation. The structure of the narrow cold-frontal rainband is characterised by young, vigorous convective cloud elements. The strong updraft, with high liquid water content and growth of precipitation particles by the accretion of liquid water contrasts sharply with the stratiform cloud and precipitation in which it is embedded (see e.g. *Houze*, 1981).

There are many studies confirming the generality of this classification of rainbands across the continental United States and in other regions of the world. Table 1 gives the cloud base height, cloud top height, rainband width, melting level and season for wide rainbands observed in the United Kingdom, (*Browning* and *Monk*, 1982), the USA (*Hobbs et al.*, 1980), Russia (*Belyakov et al.*, 1984, *Berukova* and *Shmeter*, 1989 and *Shmeter*, 1992), Finland (*Saarikivi*, 1989), China (*Gao and Zhang*, 1988 and *You and Liu*, 1989), Israel (*Gagin*, 1981) and Australia (*Ryan* and *Wilson*, 1985).

This limited survey suggests that the classification of the various types of rainbands appears to be valid over a wide range of environmental conditions. It is clear that with almost all the cloud systems, rainbands and embedded convection are a characteristic of the precipitating middle-level layer clouds. The wide mesoscale rainbands (warm frontal, warm sector and wide cold frontal rainbands) are typically 50-

Table1: The Characteristics of Wide Rainbands

Location/Reference	Cloud Top	Cloud Base	Width	0°C	References
UK	7–8km(all)	1–3km	60–150km	—	Browning and Monk, 1982
USA	~7km	~1km	50–75km	3km	Hobbs et al., 1980
	9–6km	<1km	35–70km	—	Elliott and Hovind,1964
Russia	4–6km	~1km	~50–100km	—	Belyakov et al., 1984
Finland	6.5km (summer)	—	35km	—	Saavikiv, 1983
	3.0km (winter)	—	18km	< CB	
China	1–10km (winter)	1–3km	20–40km	< CB	You and Liu, 1981
Israel	3–6km (winter)	—	20–120km	—	Gagin, 1981
Australia	4–6km (summer)	1–3km	~40km	3km	Ryan and Wilson, 1985

Table2: The Characteristics of Non Ice Enhancing Stratiform Clouds

Location/Reference	Cloud Type	Cloud Depth	n	n>20μm	d	ice	enhancement
USA: Hobbs and Rango, 1985	Ac NIPAC	<0.5 km	50–1000	0	7 μm	no	no
	AC IPAC	0.5–1.0 km	25–300	yes	10 μm	yes	no (1wc~0.1)
USA: Cooper and Vali, 1981	Cap Cloud	~0.5 km	200–600		8 μm	yes	no (1wc<1.0)
Russia: Borovikov et al. 1963	Ac	0.2–0.7	–		5–7 μm	no	no
Israel: Gagin and Neumann, 1974	Cu	3–4 km	500–1200	no	9 μm	yes	no (0.5–2.0)
Australia: King et al., 1979	Ac and As	~0.5 km	500	–	10 μm	<0.11	no

Table 3: The Characteristics of Ice Enhancing Stratiform Clouds

Location/Reference	Cloud Type	Cloud Depth	n	n>20μm	d	ice	enhancement
USA: Hobbs and Rango, 1985	Ac IPAC	0.2–0.6 km	80–500	yes	–	yes	yes (1wc>0.3)
	As/Ns	0.4–1.1 km	60–600	yes	–	yes	yes (1wc>0.3)
USA: Cooper and Saunders, 1980 (Orographic Uplift)	As/Ns	Layered	200–400	yes	20 μm	yes	yes (0.2–0.5)
Russia: Borovikov et al. 1963	Ns	several km	–	yes	7–8 μm	yes	–
China: Gao and Zhang, 1988	As (seeder)	1 km	–	no	8–18 μm	yes	no (0.1–0.5)
	St (feeder)	1 km	–	yes	4–8 μm	yes	yes (0.0–1.2)
Spain: Vali, 1989	Ac, As, Ns	1.2–6.0	200	–	–	yes	yes (0.0–0.3)
Morroco: Baddour and Rasmunsen, 1989	Ac, As, Ns	3 km	–	–	–	yes	no (0.0–0.3)
Australia: King, 1980 (No Orographic Uplift)	As/Ns	Layered	100	no	6 μm	yes	yes (1wc<0.1)
Australia: Jensen, 1993 (Orographic Uplift)	As/Ns	Layered	150–300	yes	10 μm	yes	yes (0.1–0.6)
South Africa: Bruinjes, 1988	As/Ns	Layered	100	–	20 μm	yes	yes

100 km in width, while wide mesoscale snowbands may be somewhat narrower at 20-30 km (Table 1). The narrow cold frontal rainbands bands are about 5-25 km in width.

Rainband structures have been identified during the WMO Precipitation Enhancement Project (PEP) in Spain by *Goyer* and *Cunningham* (1981), in Morocco by *Baddour et al.* (1989), in Japan in warm frontal rainbands (*Murakami et al.*, 1988 and *Murakami et al.*, 1992) and in warm sector rainbands (*Nozumi* and *Arakawa*, 1968) and South Africa (*Bruintjes*, 1988). In each of these examples the rainband width was not specified.

How important are rainbands for numerical weather prediction and for climate studies and do they need to be represented in NWP and climate models? The accurate forecasting of precipitation depends both on forecasting the timing and the intensity of rainbands. In climate studies the precipitation efficiency of precipitation events depends on both the intensity of rainbands and on the evaporation in the subcloud layer. However, from the modelling point of view it is clear that the rainband structure is resolvable only in high resolution cloud and mesoscale models and that general circulation models are unable to resolve the rainbands. This implies that rainbands need to be parameterised. This is no simple task because while the observations show that rainbands form as either wide or narrow rainbands, there are several theories proposed to explain the instabilities generating the rainbands (see *Stewart*, 1994).

To correctly model or parameterise the life cycle and the intensity of the precipitation in rainbands, it is also necessary to take account of the cloud scale processes forced by the interaction between the subcloud layer, the melting level and the underlying topography.

4.2 The Role of the subcloud layer, the melting level and topography

Regional differences in the both surface weather and the mesoscale structure of frontal layer cloud systems are forced by changes in the temperature and humidity of the subcloud and underlying surface layer, the height of the melting layer above the ground and the orographic forcing in the form of coastal convergence and topographic lifting by hills and mountains. Finally local circulations, such as land and sea breezes, can modulate the approach of fronts in coastal regions

4. 2.1 *Subcloud layer evaporation.*

On the west coast of the United States both warm sector and wide cold frontal rainbands have cloud base at about one kilometer when the low-level air beneath the warm conveyor belt is cool and moist (see e.g. *Hobbs et al.*, 1980). Consequently stratiform clouds are deep. The melting level is typically 1-2 km above cloud base. When cloud forms beneath the warm conveyor belt, the layer clouds have very high precipitation efficiencies and subcloud evaporation is minimised. For example, *Houze et al.* (1981)

estimates the efficiency of a warm frontal rainband over Washington State to be between 60-90% efficient. However, earlier *Hobbs et al.* (1980) found that two wide cold frontal bands characterised by the "seeder-feeder" process had efficiencies of 80% and 20%. In the latter case the efficiency was strongly affected by low-level evaporation.

A characteristic of the katafronts (see e.g. *Browning* and *Monk*, 1982) is that the warm conveyor belt is ascending over a cold conveyor belt that is both cold and dry. The warm frontal precipitation falling from the warm conveyor belt evaporates as it falls into the dry subcloud air (Figure 3).

During summer in south-eastern Australia the sub-cloud layer beneath the warm conveyor belt is cloud free. As shown by *Ryan et al.* (1989) this hot northerly airstream in the sub-cloud layer can extend several 100 km off-shore before being modified. Relative humidities of less than 30% in the prefrontal sub-cloud layer air are common (see *Ryan* and *Wilson*, 1985). Cloud base is typically 3-4 km over the land for the warm sector rainband with the freezing level near 3 km. The subcloud layer is cooled and moistened by the virga falling from the mid-level layer cloud. Using a nonhydrostatic cloud model, *Ryan et al.* (1990) showed that the evaporative cooling generated by virga falling from the middle-level cloud was sufficient to generate the observed squall line in the cloud layer. By contrast in southern Australia both during winter and spring the prefrontal air- stream associated with cutoff lows is both cooler and moister and cloud base is more typically 1-2 km. Under these conditions it is not uncommon for a cloud system to comprise several layers each about one kilometer thick and separated by a few hundred meters. (*Jensen*, pers. comm.).

4.2.2 *The dynamical forcing by the melting layer*

In many areas of the world, the precipitation at the surface characteristically associated with winter storms varies in form from snow to freezing rain and ice pellets and then finally to rain. *Stewart* (1992) points out that a favorable location for the precipitation-type transition is in the vicinity of coastlines. In both Canada and Scandinavia extremely heavy precipitation has been observed and characterised by snow over the land, where temperatures were below freezing, a mixture of rain and snow along the coast and rain over the ocean. Stewart suggests the scale of the transition is about 25 km and that the transition region has a dynamical structure that leads to mesoscale circulations in close proximity to the transition region thereby modifying the location and intensity of the precipitation.

4 2.3 *Orographic Forcing of Layer Clouds*

There are many examples from around the world on the impact of orography on cold fronts. It is beyond the scope of this paper to give a detailed review of fronts and orography. An extensive review of both observations and theory is given by *Egger* and *Hoinka* (1992). Egger and Hoinka list several observational programs that have studied the impact of orography on fronts. The majority of these observational programs focused on the dynamical structure of the fronts and relatively few studies have concentrated on the cloud structure.

It is clear that orographic forcing considerably modulates the precipitation associated with layer clouds. The direct effects are generally to increase the upward flow of air on the windward side which in turn enhances the formation of clouds and precipitation. On the leeward side there is descent that suppresses the formation of clouds and precipitation. The forced ascent over the hill or mountain causes the formation of "feeder" clouds. Above the "feeder" clouds are "seeder clouds forced by the large scale ascent. The precipitation generated in the "seeder" clouds is enhanced by the water rich environment of the "feeder" clouds leading to increased rainfall. The seeder- feeder mechanism is most effective in the warm sector of an extratropical cyclone (see e.g. *Hobbs*, 1978).

Large mountain ranges are an important constraint on the development of the mid-level cloud systems on the synoptic scale. For example, in the USA, development of cyclonic storms is constrained by mountain ranges. Upstream of the Rocky Mountains, the orographic barrier partially cuts off the moisture in the warm conveyor belt coming from the south into the frontal system, while on the leeside, snowfall along Colorado's Front Range is linked to cold air damming. Frontal over-running of the trapped cold air by a moist south-easterly flow causes heavy precipitation to occur (see e.g. *Cotton* and *Anthes*, 1989).

From the modelling perspective the effect of orography on cloud and precipitation is important and there is a continuing need for both process studies and model validation. The review by *Egger* and *Hoinka* (1992), point to several data sets in the both Europe and the USA that should be suitable to validate the generation of middle level cloud in the vicinity of mountain ranges by models. In particular, there seems to be very complete data sets associated with CYCLES, GALE and the German contribution to Front87. However there remains a need for high quality data sets from other regions of the world.

4 2.4 *Coastal and Diurnal Forcing*

Coastal effects modulate frontal cloud cover and this has been observed in different geographic locations. For example, *Dodge* and *Burpee* (1993) showed that on the mid- Atlantic east coast of the United States rainband activity was modulated by the diurnal cycle and showed land-ocean differences in the structure of the rainbands. In particular, there was a greater occurrence at night of both strong echoes (convective cloud) and weak echoes (stratiform cloud) 100 - 200 km off shore and parallel to the coast. A second example is in Australia where *Garratt* (1987) showed that coastal effects, such as the development of a sea breeze, lead to a tendency for a high percentage of fronts to cross the coastline in the afternoon and early evening hours.

4 2.5 *Summary*

While observations on the synoptic scale support the conceptual model of precipitating layer clouds, and there is generality in the currently accepted classification of rain bands, the nature of the bands and the layer clouds may be modulated by processes occurring in the subcloud layer. From a climate modelling perspective the rainband circulations and the associated evaporative cooling, melting layer effects, orographic lifting, coastal contrasts and diurnal effects may not be resolved by the model and hence require parameterisation. These effects are important climatologically and should not be ignored as they markedly affect the precipitation efficiencies of the layer cloud systems.

5. MICROPHYSICAL CHARACTERISTICS OF PRECIPITATING LAYER CLOUDS

In climate simulations the microphysical structure of the clouds is important. Water clouds are radiatively different from ice clouds. To calculate the radiative response of a cloud it is not only necessary to know the phase of the water in the cloud, it is also necessary to characterise the particle size distribution of the clouds, the ice crystal habit and the thickness of the clouds. In addition to determining the radiative feedback of the clouds, the microphysics determines the precipitation efficiency of the precipitating cloud system. In mid-latitudes, precipitating layer clouds usually extend above the freezing level. It is convenient therefore to classify the layer clouds into non-ice producing stratiform clouds, non-ice enhancing stratiform clouds and ice enhancing stratiform clouds as defined by *Hobbs* and *Rangno* (1985) and described in the next section.

5.1 **Ice crystal formation in precipitating layer clouds**

5. 1.1 *Non-Ice Producing and Non-Ice Enhancing Stratiform Clouds*

Hobbs and *Rangno* (1985) described the cloud microphysical structure of Altocumulus, Ac, Altostratus, As, and Nimbostratus, Ns. "Non-ice producing Ac", NIPAC, contained no ice crystals or relatively low concentrations of ice crystals, (<10 per litre). These clouds were generally thin and often newly formed. Those clouds containing ice crystals were often thicker and longer lived and they refer to them as "ice producing Ac", IPAC, clouds A feature of the IPAC clouds is that there tends to be very little ice enhancement. Ice enhancement or ice multiplication is said to have taken place if the observed numbers of ice crystals substantially exceed the number of ice crystals that would have formed on the observed background ice nucleus population. In all but one of the examples given by Hobbs and Rangno the altocumulus clouds having ice multiplication had significant liquid water contents near cloud top exceeding 0.3 g m^{-3} .

There are many observations to show that stratiform clouds less than one kilometer deep may have either very few ice crystals or no ice crystals (Table 2). *Heymsfield* (1977) found that jet stream cloud tended to be uniform with some liquid water as low as -35°C. More recently, *Heymsfield* et al. (1991) examined two thin altocumulus clouds at about -30°C. They point out that the absence of ice crystals implies a dearth of ice nuclei and this is contrary to the assumption that ice crystal concentrations increase with decreasing temperature. They hypothesised that since ice nuclei concentrations in cirrus cloud are also low while ice crystals are abundant, there is support for the idea that ice crystals produced at temperatures below -40°C originate through the homogeneous freezing of water droplets.

The only deep clouds reported to have ice crystal numbers close to that predicted by the observed ice nucleus spectrum are in Israel (*Gagin* and *Neumann*, 1974). However, recently *Rangno* and *Hobbs* (1988) have questioned these findings as they appear to be inconsistent with similar cloud observations found in the United States. This criticism is supported by recent measurements made by *Levin* (1994) in Israel. Levin found ice crystals in these clouds at concentrations of up to 3 orders of magnitude above that expected on the basis of ice nuclei measurements.

5. 1.2 *Ice Enhancing Stratiform Clouds*

Hobbs and *Rangno* (1985) suggest that the difference between the non-ice producing stratiform clouds and high concentrations of ice crystals found in the deeper stratiform clouds is the breadth of the water droplet spectra. In particular, they noted that the IPAC clouds had droplets with diameters > 20 μm in concentrations of more than 10 cm . Table 3 shows observations from USA, Russia, China, Spain and Australia that support their observations. In all cases the clouds glaciated and, with the exception of the seeder clouds in China (*Gao* and *Zhang*, 1988) and Western Victoria in Australia (*King*, 1980), the cloud droplet spectra had observable concentrations of droplets greater than 20 μm in diameter.

Generally, the deeper systems were largely glaciated cloud and accompanied by embedded convection. This is true of most of the observations shown in Table 3. A feature of several recent studies is that while these clouds are largely glaciated, water droplets are often observed near the top of the clouds. Figure 4 shows an example from *Paltridge* et al. (1986). Calculations by *Rauber* and *Tokay* (1991) suggest that the liquid layers occur when there is an imbalance between the condensate supply rate and the bulk ice crystal mass growth rates. They demonstrated that the small ice crystals at cloud top allow the existence of a liquid water layer in weak updrafts over a wide range of temperatures and ice crystal concentrations. Rauber and Tolkay argue that entrainment, shear and radiative cooling processes near cloud top are able to induce the instabilities to generate the updraft regions.

The sensitivity of radiation calculations in GCM and limited area models to the presence of these liquid water layers is largely untested. Analysis of additional data sets are needed to verify the generality of these observations supported by modelling studies to calculate the lifetime and radiative response of the liquid water layers in the clouds. However calculations by *Platt* (pers. comm.) suggest that for the short wave

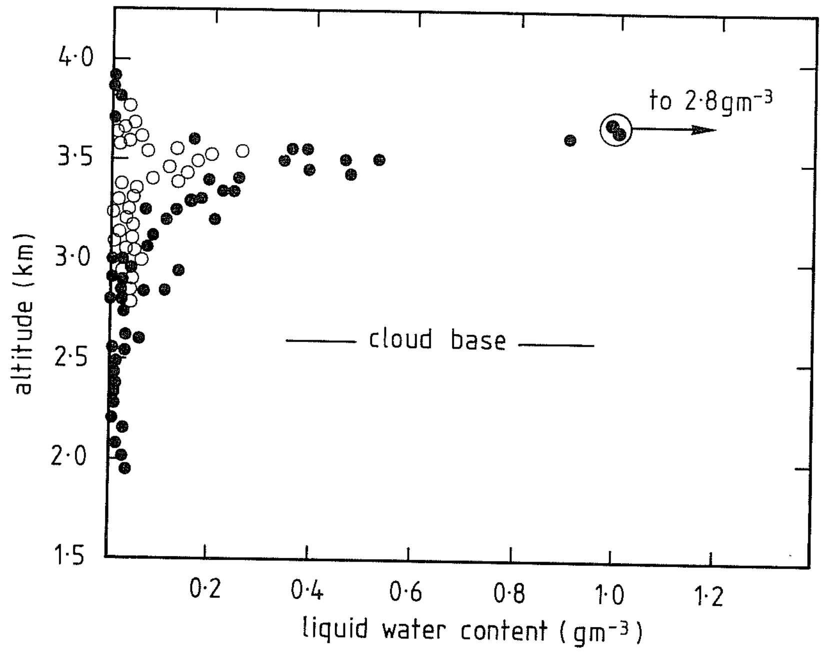

Fig. 4a Liquid water content of a mixed phase altostratus cloud showing the high liquid water contents neat cloud top for two different profiles of the same cloud are shown as solid and open circles (from *Paltridge* et al. 1986).

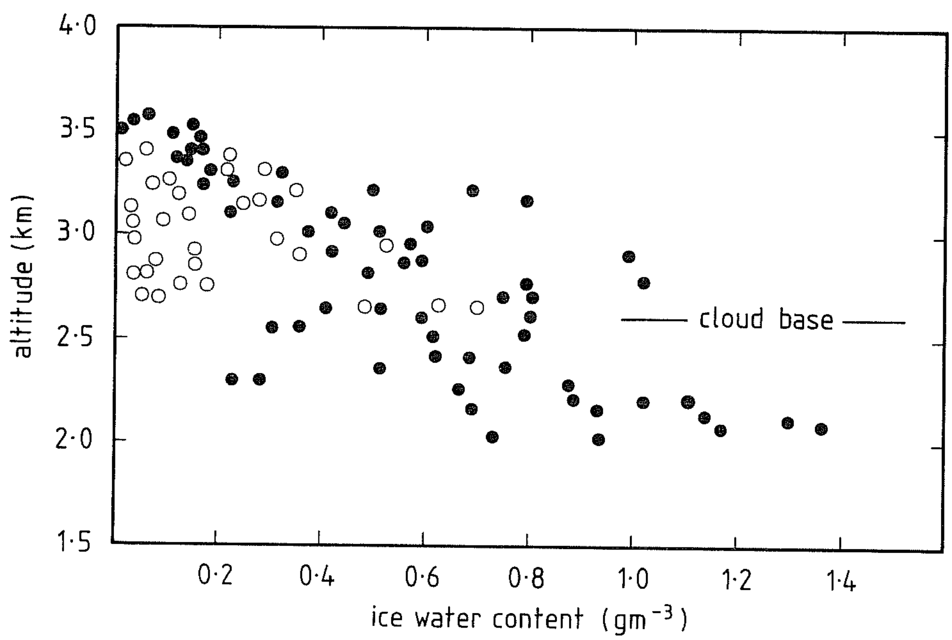

Fig. 4b Ice water content profiles through the cloud shown in Figure 4a. (from *Paltridge* et al. 1986)

albedo, a 100 m layer of small water drops is equivalent to 1 km of ice crystals. Consequently if these liquid layers are ubiquitous they may be radiatively significant.

In deep layer clouds, observations suggest that when cloud tops are warmer than -15°C liquid water contents exceeding 0.2 g m^{-3} are found frequently in cloud (see e.g. *Borovikov* et al., 1963) This is particularly true when the layer clouds are modified by orography. Relatively high liquid water contents are found in the regions of embedded convection. Away from regions of embedded convection the cloud water contents are small and the clouds are glaciated.

Generally, more observations of the ice water content of layer clouds are required to both test theories of ice crystal generation and to develop parameterisations for GCM and limited area models. There are two important gaps where there is a severe lack of data. First, there is a dearth of information on layer clouds over mid-oceanic areas and secondly, observations over continental areas in non-mountainous regions are comparatively rare.

5. 2. Ice crystal numbers in clouds

Hobbs and *Rangno* (1985) made an extensive study of ice crystal numbers in stratiform cloud that formed over Washington State, Southern Utah and the Pacific Ocean. Their extensive study included altocumulus, altostratus, nimbostratus, stratocumulus and stratus. In this paper only the observations relating to the altostratus and nimbostratus cloud systems will be discussed (see Table 3 of their paper). Hobbs and Rangno conclude that there was no significant correlation between the maximum number of ice crystals and cloud top temperature. However, their results suggest that there is a relation between the number of ice crystals and the width of the droplet spectrum, namely:

$$I = \left(D_T / 19.4\right)^{6.6} \tag{1}$$

where I is in l^{-1} and D_T is the diameter in æm that exceeds a cumulative concentration of 3 cm^{-3}. The onset of ice particles in the stratiform cloud with tops colder than -6°C commences with the development of larger droplets and is independent of the cloud top temperature. The cloud drop spectra of the NIPAC are narrow and the concentration of droplets greater than 20μm is less than 1 cm^{-3}, while the high ice producers have concentrations of droplets with diameters greater than 20μm exceeding 10cm^{-3}. In other regions of the world, there is some evidence that in precipitating stratiform cloud systems there may be a relationship between cloud top temperature and the number of ice crystals near cloud top. These data come from observations in south eastern Australia (*King*, 1982 and *Mossop*, 1968), Colorado, USA (*Cooper* and *Saunders*, 1980), northern Spain (*Vali* et al., 1988) and Israel (*Gagin*, 1981), and are summarised in Figure 5. Also shown in Figure 5 are the results from Table 3 of *Hobbs* and *Rangno*

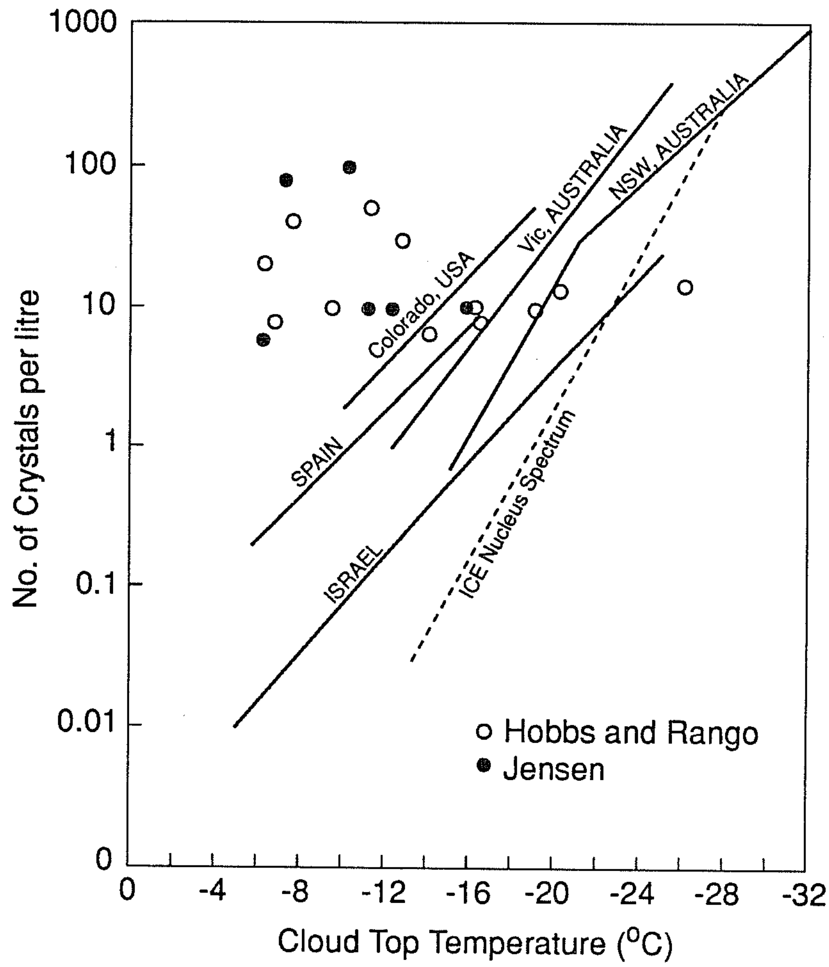

Fig. 5 Observations of ice crystal concentrations (no. per litre) as a function of cloud top temperature. The ice nucleus spectrum is from *Fletcher* (1962), the Australian data from *Mossop* (1968) and *King* (1980), the Spanish data from *Vali* et al. (1988), and the Colorado data from *Cooper* and *Saunders* (1980) are shown as linear fits to the data. Observations from *Hobbs* and *Rangno* (1985) over the Pacific North West of the USA are shown as open dots and observations from the Australian Winter Storms Experiment (*Jensen* pers. comm.) are shown as closed dots.

(1985) and observations from the Australian Winter Storms Experiment (AWSE) (*Jensen*, pers. comm.). The ice crystal number dependence on cloud top temperature for the Australian Winter Storms Experiment is more consistent with that of *Hobbs* and *Rangno* (1985) than those found by *King* (1982) in Western Victoria and *Mossop* (1968) in New South Wales. The differences in the droplet spectra found by *King* (1982) over Western Victoria and by *Jensen* (pers. comm.) near the Australian Great Dividing Range are shown in Table 3. The Western Victorian observations show very few droplets greater than 20μm, a mean diameter of about 6μm and low liquid water contents in the cloud of less than 0.1 gm^{-3} . By contrast Jensen's results show droplets greater than 20μm, a mean droplet size diameter of 10μm and liquid water contents exceeding 0.1 $g\ m^{-3}$. The ice crystals form in synoptically similar systems with the main difference being the effect of the Dividing Range on the cloud system. In the observing region for the AWSE the orography has a maximum peak of 1.2 km. This modest topographic barrier appears to change substantially the microphysical structure of the clouds. Since most observations of the ice crystal numbers in precipitating layer clouds have been made near regions of steep topography there is a need to confirm their representativeness away from significant topography by comparing them with new observations that are either over the sea or over flat land.

The observations by *Bruintjes* (1988) in South Africa are interesting because above the melting level in the prefrontal and warm frontal rainbands supercooled liquid water is encountered on only 5% of occasions. This is comparable with the observations in Australia over western Victoria (*King*, 1982). Near cloud base the droplet concentration is typically 100 cm^{-3} and the mean diameter of the droplets is 20μm. On the other hand the prefrontal cold surge was substantially more convective than the prefrontal and warm frontal rainbands. In the convective elements liquid water was present in new towers and ice crystal concentrations between 10 l^{-1} and 100 l^{-1} were found. However in regions where the convective activity was weak the ice particle concentrations were in the range 10-20 l^{-1}. The cloud top temperature is not given in the paper, however, the cloud top was estimated to be about 6 km. In summary there is a need to find observations that will test the generality of the empirical relation found by *Hobbs* and *Rangno* (1985) as well as to determine under what conditions there is a relationship between cloud top temperatures and ice crystal concentrations.

5.3 Ice Crystal spectra

As well as ice crystal numbers the second parameter required to calculate the precipitation and radiative response of precipitating layer clouds is the shape of the particle spectrum. *Platt* (1995) examined the particle size spectra from precipitating clouds provided by Dr. A. Heymsfield. Some of the data from non-precipitating cirrus had been used previously by *Heymsfield* and *Platt* (1984) in a study of the particle size spectrum of cirrus clouds.

These authors argue that the particle size spectra in the range from 10 μm to about 100 μm obey a power law of the form:

$$N(D) = aD^b \tag{2}$$

where N(D) is the number of ice crystal (m) and D is the largest dimension of the ice crystal in μm. Heymsfield and Platt found that "b" varied from -2.2 to -3.8 over the temperature range -25°C to -60°C.

Platt (1995) found similar power laws for precipitating clouds but, for particle sizes in excess of 100-200μm, then the size distribution follows more closely that of a Marshall Palmer distribution,

$$N(D) = N_0 \, exp(-\lambda D). \tag{3}$$

Figure 6 shows an example from *Platt* (1995) showing both the power law segment and the exponential segment for particles collected between -25° C and -30° C in the same cloud. Similar shaped spectra have been reported by *Paltridge* et al. (1986) in Australia, *Houze* et al. (1979) in the USA and *You et al.* (1990) in China, although the power law component of the parameterisation was not derived.

Many NWP and climate models parameterise the ice phase in the microphysics in terms of precipitation and cloud sized ice crystals. Based on Platt's observations, there is a good case for parameterising precipitation sized particles in terms of the Marshall Palmer distribution, while parameterising the smaller cloud sized particles using the power law. The problem with this formulation is that there is no physically based "auto conversion" parameterisation that allows for generation of precipitation sized ice particles from the ice cloud. This is clearly a problem that needs to be addressed.

The Marshall Palmer component of the distribution for precipitating clouds has been studied extensively in a number of regions of the world and a consistent picture has evolved. Figure 7 shows examples of the exponent λ as a function of temperature for Washington State USA (*Houze et al.* 1979), California (*Stewart* et al., 1984 and *Gordon* and *Marwitz*, 1986), Australia (*Jensen*, pers. comm.), China (*You* and *Liu*, 1989), Europe (*Bennetts* and *Ryder*, 1984 and *Marecal*, 1993) and Figure 8 shows examples of the value of N_O as a function of temperature for Washington State, California, Australia and Europe. In their analysis, *Gordon* and *Marwitz* (1986) identified a change in the slope of λ with particular ice crystal growth regimes. Apart from these subtle changes, λ is a simple function of temperature in all of the data sets and therefore can be parameterised independent of the geographic region. The functional relation between N_0 and temperature is less systematic than that for λ with local variations being evident.

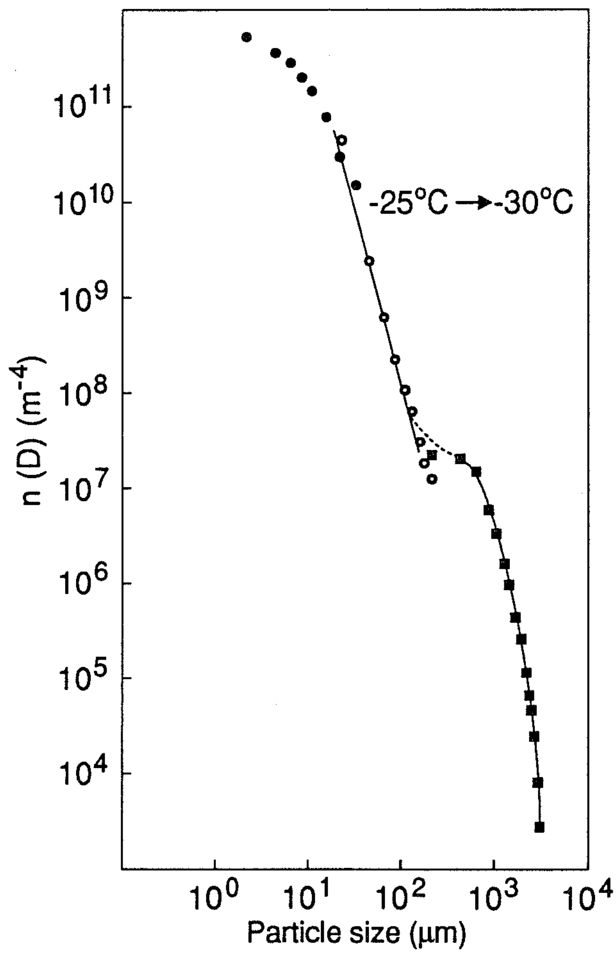

Fig. 6 Observations of the observed particle concentrations from a layer cloud in the *Heymsfield* and *Platt* (1984) data set (courtesy C.M.R. Platt).

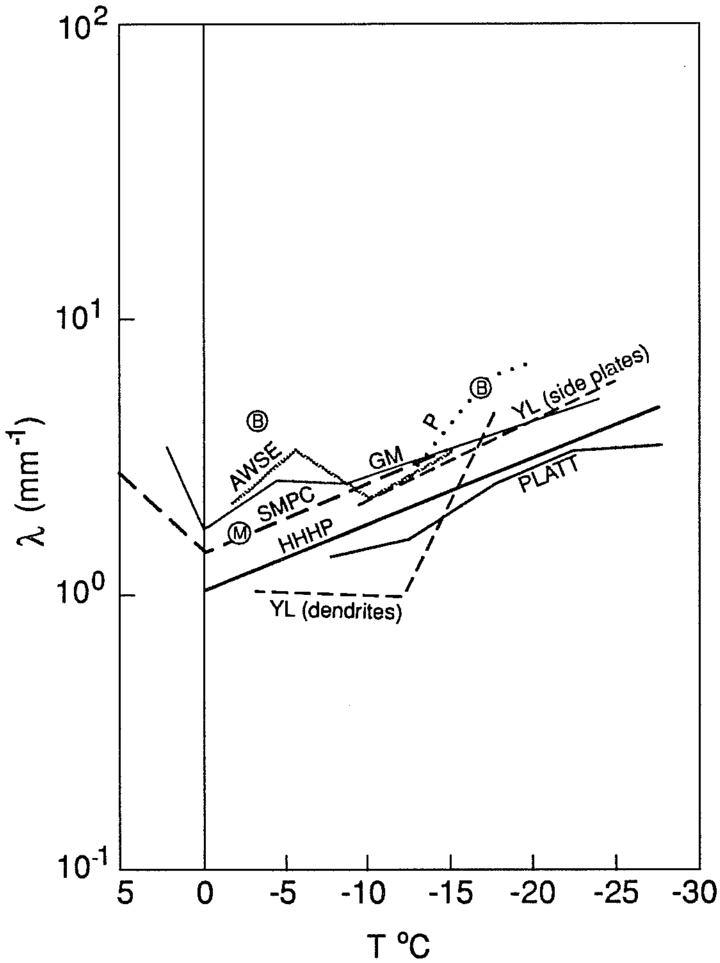

Fig. 7 The observed exponent λ (m^{-1}) for the Marshall Palmer distribution in clouds as a function of temperature. HHHP is from Washington State (*Houze* et al. 1979), SMPC is from California (*Stewart* et al, 1984) as is GM (*Gordon* and *Marwitz*, 1986), Platt from *Heymsfield* and *Platt* (1984), AWSE from Australia (*Jensen*, pers. comm.), YL from China (*You* and *Liu* (1989)), and from Europe, *Bennetts* and *Ryder* (1984), shown as (B) and *Marecal*, (1993), shown as (M).

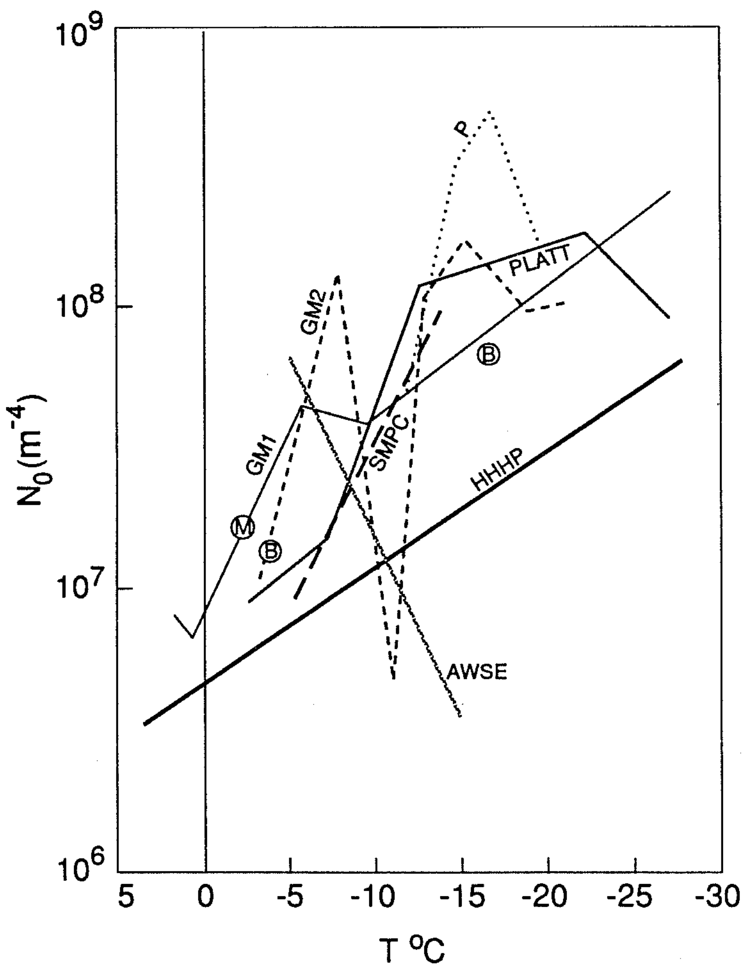

Fig. 8 The observed constant N_0 (m^{-3}) for the Marshall Palmer distribution in clouds as a function of temperature. The symbols are as in Figure 8.

6. SUMMARY

The results of the study quantitatively illustrate that the synoptic and mesoscale characteristics of precipitating layer clouds around the world are not strongly dependent on the geographic location. However, despite the large number of field experiments undertaken in recent years, it is clear that these conclusions are biased to observations taken near mountainous terrain or near the coast. There are almost no observations over the open oceans.

The simple conceptual conveyor belt model of frontal systems gives a clear understanding of the flows generating the layer cloud systems on the synoptic scale. Both global models and limited area models have the resolution to diagnose the flows in the conceptual model.

The efficiency and intensity of rainbands generated by mid-latitude cloud systems may be modulated by evaporation and sublimation in the subcloud layer and by orographic and coastal forcing. It is doubtful if the current generation of general circulation models has the physics or resolution to be able to model these processes.

The microphysical structure of the precipitating clouds is dominated by the ice phase in the clouds. The life time of liquid water above the melting layer is small away from regions of enhanced convection and orographic uplift. Observations suggest that for ice particles with sizes between 10μm and 100μm the distribution is best described by a power law and for sizes above 200μm the ice crystal size distribution is best described by a Marshall Palmer distribution. The survey would suggest that fundamental knowledge needed to parameterise the number of ice crystals in clouds is still partially incomplete.

It is clear that a systematic data base for precipitating layer clouds is required and this is a priority for the GEWEX Cloud System Study Working Group III (*Browning*, 1994). This paper has identified two uses for such a data base, namely to provide data for validating cloud climatologies generated by climate models and to provide data for process studies. Work is in progress on defining the specifications for the data base. This paper has identified some of the variables required by the data base but more importantly has demonstrated the incomplete nature of most data sets.

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Wexler, R. and D. Atlas, 1959: Precipitation generating cells. J. Met., **16**, 327-332.

9. APPENDIX 1

CYCLES	CYCLonic Extratropical Storms Project Pacific North–West United States: *Hobbs et al., 1980*
CAO	Central Aerological Laboratory Russia, 1977–1979: *Postnov, 1983*
CFRP	Cold Fronts Research Program South–Eastern Australia, 1980–1984: *Ryan et al., 1985*
CASP I & II	Canadian Atlantic Storms Program, 1986 and 1992 Western Atlantic: *Stewart et al., 1987 and Stewart, 1991*
ERICA	Experiment on Rapidly Intensifying Cyclones over the Atlantic, Western Atlantic, 1986–1991: *Hadlock and Kreitzberg, 1988*
GALE	Genesis of Atlantic Lows Experiment East Coast United States, 1986: *Dirks et al., 1988*
BASE	Beaufort and Artic Storms Experiment *Beaufort and Artic Storms Experiment, 1993*
FRONTS87	Eastern Atlantic, 1987: *Clough and Testud, 1988*

10.APPENDIX 2

ISRAEL I & II	Israel, 1961–1967 and 1967–1975: *Gagin, 1981*
CASCADE	Pacific North–West United States, 1972–1973: *Hobbs, 1975*
CRBPP	Colorado River Basin Pilot Project Colorado, United States 1974–1975: *Marwitz, 1980*
SCPP	Sierra Cooperative Pilot Project Nevada, United States, 1979–1980: *Heggli et al., 1983*
WVCSE	Western Victorian Cloud Seeding Experiment South Eastern Australia, 1978–1980: *King, 1982*
COSE	Colorado Orographic Cloud Seeding Experiment, Clorado, United States, 1979–1981: *Rauber and Grant 1986*
TAS II	Tasmania II Tasmania, Australia, 1979–1982: *Shaw et al., 1984*
PEP	Precipitation Enhancement Project, 1979–1984 Villanubla, *Spain, Vali et al., 1988*
X'JIANG	Xinjiang Experiment, 1982–1984 Urumiq Region, China: *You et al., 1990*
AME	Atlas Mountains Experiment, 1984 Morocco: *Baddour and Rasmussen, 1989*
AWSE	Australian Winter Storms Experiment, 1988–1992 South–Eastern Australia: *Long and Huggins, 1992*