Microphysics: From intricacy to simplicity

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1. Introduction

This paper discusses the parametrization of cloud and precipitation microphysical processes in atmospheric numerical models. There are many different aspects to this problem, but the focus here is specifically on the representation of microphysics and the interaction with atmospheric dynamics through diabatic processes, with an emphasis on parametrization for global Numerical Weather Prediction (NWP) and climate models (general circulation models or GCMs). The parametrization of partial cloudiness in a model grid-box and sub-grid heterogeneity of vapour, condensate and hydrometeors play an important role in such models but these issues are covered by Tompkins paper in this volume and are therefore not discussed here.

First there is a brief overview of microphysical processes in the atmosphere to convey a sense of the wide range of scales involved from the molecular- to the macro-scale (Section 2). This is followed in Section 3 by a discussion of different approaches to the parametrization of these microphysical processes and some general issues relevant for parametrization development. The interaction of the microphysics with dynamics through latent heating and cooling of the atmosphere is then reviewed and discussed in Section 4 with emphasis on a need to understand the interactions between different parts of a model to make progress in improving the accuracy of model forecasts. The concluding section provides a summary and suggests areas for future development of microphysical parametrization in atmospheric numerical models.

2. Intricacy and Complexity: Microphysics in the Atmosphere

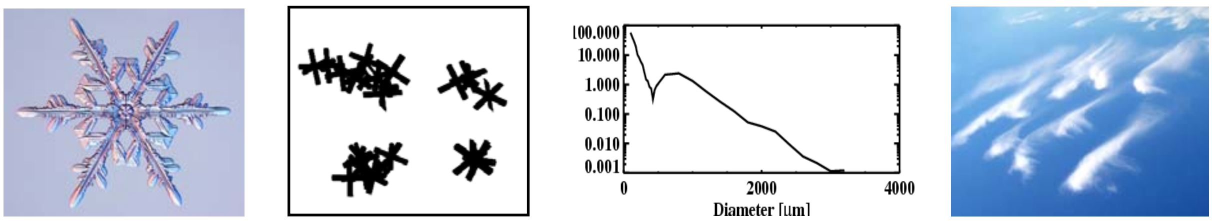
The subject of microphysics in the atmosphere is a large topic, with much research from the mid-1900s to the present day and only a few words are provided here to highlight the different scales involved. There are many books and a wide-ranging literature that provide the detail of our current understanding of cloud and precipitation microphysics to which I refer the reader for further in-depth information (e.g. Pruppacher and Klett, 1996; Rogers and Yau, 1989).

In order to understand microphysics we need to understand what happens at the molecular scale. Phase transitions are characterised by the presence of free energy barriers at the molecular scale, which must be overcome to form cloud droplets and ice crystals (called nucleation). Nucleation of water from vapour, ice from vapour and ice from water are either homogeneous (i.e. a 'pure' system) or heterogeneous (i.e. a system containing impurities). Homogeneous nucleation for liquid water droplets requires very high supersaturations to overcome the surface tension or free energy barrier. In fact such vapour pressure gradients are not observed in the atmosphere, as water droplets are formed at much lower supersaturations through the heterogeneous nucleation process of condensation on hygroscopic cloud condensation nuclei (CCN). There are generally abundant CCN present in the atmosphere from natural and anthropogenic sources such as dust, sea spray, forest fires and combustion, but the concentrations vary spatially and temporally with usually higher source values over land and lower over the oceans. Nucleation of ice crystals can also proceed via a homogeneous or heterogeneous process, but ice nuclei (IN) are scarcer, have different activation and significant supersaturations with respect to ice can be supported.

Once a water droplet or ice crystal is formed, it can grow further through the diffusion of water molecules from the vapour phase on to its surface. Particularly for the ice phase, this process is complex and leads to the wide variety of ice crystal shapes that are present in the atmosphere; the mode of growth depending on both the temperature and supersaturation and determining the shape (habit) of the growing crystal.

Nucleation of particles, condensation and diffusional growth, evaporation and sublimation all depend on the properties of molecules and their interactions and determine the formation and growth/decay of “cloud”. However, diffusion growth theory alone cannot explain the rapid conversion of cloud water droplets to raindrops, and ice crystals to large snowflakes, in the times observed in real clouds. These precipitation processes can be explained by inter-particle collection processes (often called collision-coalescence or accretion for water drops, aggregation for ice crystals and riming for ice crystals capturing water drops). Initially the growth of a population of droplets or particles by differential diffusion leads to a broadening of the spectrum of fall velocities and increases the likelihood that two drops will collide with each other. The processes involved in the initial broadening of the particle spectrum are still debated but local evaporation in a heterogeneous water vapour field created by small scale turbulent motions is a possibility as well as giant condensation nuclei. Once the particle size (and fall velocity spectrum) is broad enough, collection then takes over as the dominant process leading to particles large enough to precipitate.

Saturation, defined as the equilibrium situation in which the rates of evaporation and condensation (or ice deposition and sublimation) are equal, is clearly a key concept in atmospheric microphysics. Whether the air is supersaturated or subsaturated with respect to water or to ice is primarily determined by gradients of temperature and pressure and atmospheric motion, with ascent producing supersaturation and descent creating subsaturation. It is atmospheric dynamics, from small-scale turbulent motions to large-scale air mass movements and the feedbacks through the latent heating and cooling of microphysical processes, that create the wide variation in cloud and precipitation, from individual cumulus clouds to mesoscale convective systems, stratocumulus to frontal mid-latitude cyclones. Figure 1 is a simple illustration of the wide range of scales associated with microphysics in the atmosphere.



(a) (b) (c) (d)

Figure 1. Microphysics across a range of scales: (a) a dendrite ice crystal formed by diffusional growth (from SnowCrystals.com) $O(10^{-6}$m), (b) aggregation of a number of bullet rosettes ice particles (from C. Westbrook) $O(10^{-3}$ m), (c) an ice particle size distribution (diameter vs. concentration) (from P. Field) in a volume of air of O(1 m) and (d) a large ensemble of ice particles $O(10^{3}$ m), i.e. cirrus clouds!

3. Simplicity and Approximation: Parametrizing Microphysics

Microphysical processes are complex micro-scale phenomena and the collective effect must be greatly simplified and approximated in numerical models used for NWP and climate prediction. The complexity of the parametrization depends on the available computing power, the particular application of the numerical model, and the degree of our understanding! The parametrization in a model designed for studying small-scale convective processes may be very different to a parametrization designed for a global climate model.

First we may ask the basic question of why we need to represent microphysics in a GCM. There are three primary reasons:

- Water cycle: Representation of clouds and precipitation, their spatial and temporal variations and modification of the surface hydrology.
- Radiative impacts: Modification of the radiative fluxes in the atmosphere and at the Earth's surface through absorption and emissivity.
- Dynamical impacts: Primarily through the heating and cooling of the atmosphere related to phase changes of water.

In addition there may also be other reasons to represent microphysics more accurately such as model validation or assimilation with remote sensing observations of cloud.

This section discusses the different complexities of microphysical parametrizations used in numerical models, the representation of some of the principle microphysical processes and a few thoughts on microphysical parametrization development.

3.1. Hierarchy of Parametrization Schemes

There are many microphysical parametrization schemes described in the literature and the primary differences between schemes are the number of variables used to represent the different phases of water, the number of parameters required to describe particle size spectra, and the details of the microphysical interactions and particle characteristics.

It is convenient to place cloud and precipitation particles into a number of discrete categories. As well as water molecules in the vapour phase, liquid and solid particle types in the atmosphere can be described in terms of cloud water droplets, raindrops, pristine ice crystals of varying shape (habit), aggregates of pristine crystals, rimed ice crystals, graupel (heavily rimed ice particles) and hail. A given volume of air may contain a number of particles of different types with a wide spectrum of sizes undergoing complex collisions and phase change processes. It is the purpose of a microphysical parametrization to represent the combined effects of this complex system within some specified volume (e.g. a model grid-box).

The governing equation for the time evolution of a moist variable q_k (which could represent any of the above types of particle) is

$$\frac{\partial q_k}{\partial t} = \vec{u}\nabla q_k + \frac{\partial q_k v_k}{\partial z} + F_{q_k} + \phi_{q_k} \tag{1}$$

where q_k represents a collection of particles of phase k, for example, by a mass mixing ratio or number concentration. The first term on the right hand side represents advection by the three-dimensional wind ($\vec{u}$), the second term represents the vertical divergence of q_k due to sedimentation under the influence of gravity with an effective terminal fall speed (v_k), the third term F_{qk} represents the diffusive effects of turbulent fluxes and the last term ϕ_{qk} represents the sources and sinks and inter-particle interactions of q_k.

All microphysical models/parametrization schemes represent some or all of the terms in Eq. 1 but with different degrees of approximation. The number of particle categories that are represented is one difference between parametrizations, but another fundamental difference is the way in which the wide spectrum of particle sizes is represented. The following sub-sections discuss the particle size distribution and most of the terms on the right hand side of Eq. 1.

(a) Representing the particle size and mass distributions

Lagrangian or *particle-tracing* models (e.g. Young, 1993) follow individual particles or groups of particles with the advantage that the particle history can be traced and analysed. *Bin models* (e.g. Young 1975; Cardwell et al. 2002, Lynn et al. 2005, Morrison and Grabowski 2007) split the particle size spectrum into discrete bins and are more appropriate for microphysical modelling when the particle interactions become too complex for particle-tracing models. Each bin represents either a particular particle size or a continuous range of sizes and explicit microphysical interactions transfer particle numbers and mass between bins. This enables processes such as autoconversion (rain formation from cloud droplets) to be modelled explicitly, although there remains significant uncertainty in the parametrization of some of the interaction equations; for example, the effect of turbulence on the efficiency of inter-particle collisions. Lagrangian and Eulerian bin parametrizations are generally computationally expensive and are used for detailed microphysical modelling with the emphasis on understanding the microphysical processes that occur in the atmosphere, although increasingly they are being coupled to dynamical models (Lynn et al. 2005, Morrison et al. 2005).

A third category of models assume a simple functional form for the whole of the particle size spectrum and use equations that describe the evolution of the parameters of the size distribution function as the particles evolve. This is the form that is generally used in GCMs due to its relatively simple formulation and computational efficiency. Such schemes are often termed 'bulk' parametrization schemes as the characteristics of the size distribution are directly related to bulk quantities (i.e. grid box values) in the model. The actual variable or variables represented by q_k for a particular particle type could be one or more of the moments of the distribution (e.g. mixing ratio, number concentration, mean diameter, size spectrum slope) as long as all the parameters of the size distribution are derivable, ideally through a function that can be easily integrated. Observations of rain size spectra (Marshall and Palmer, 1948) and snow size spectra (Gunn and Marshall, 1958) show that an exponential function is a first order approximation to the shape of snow and rain particle size distributions and this form is widely used, although alternatives such as gamma and log-normal functions are also in use.

The mass distribution depends on the number density and particle mass for the range of particle sizes. The mass of all cloud water droplets and raindrops is essentially constant at 1000 kg m^{-3}, but the bulk density of ice particles varies significantly with particle size and type due to their complex geometry.

(b) Advection and sedimentation

The first two terms on the right hand side of Eq. 1 represent the advection and sedimentation of particles. The relative importance of each term is different for different particle types and in terms of parametrization is also dependent on the grid box size of the model. At one extreme, if the fall speed of the particles is zero, then only the advection term, representing the transport by the three-dimensional wind, is required. At the other extreme, if the particle or group of particles of a particular type fall from cloud to ground in less than the time it takes for advection by the horizontal wind across a grid box, then only the sedimentation term is important and horizontal advection can be neglected. For the intermediate case, both terms may be required.

Sedimentation of cloud liquid water droplets is small and often neglected, so only the advection term is required. Very small ice crystals also come under this first category, but the fall speed of ice particles increases with particle size (to around 1 m/s for unrimed particles) and generally both sedimentation and advection terms are of importance. There is a clear distinction between the size distributions and terminal fall speeds of cloud water droplets and rain drops that necessitates the separate representation of the two distinct particle types in a model. However, the conversion of ice crystals to snow (aggregates of ice crystals) does not exhibit as distinct a mode separation and so some models choose to represent all ice/snow as one variable. Rain, graupel and hail have much higher terminal fall speeds (5-10m/s), and horizontal advection is relatively small in the time it takes to fall from cloud to ground. Thus only the sedimentation term is required for these particle types unless the horizontal resolution of the model is high.

The timestep of the model is also relevant for the representation of the sedimentation term. If the particle type falls to the ground in less than a model timestep then a numerical formulation of sedimentation is not required as an equilibrium can be assumed, but if it takes many timesteps for the particle to fall through the atmosphere then a numerical formulation can be used to determine the new location at the end of the timestep. In this case the terminal fall speed of the particles becomes an important parameter defining the vertical flux divergence.

Based on the above arguments, both the model grid resolution and the model timestep determine which particle types need to be held in the model as prognostic variables (i.e. values are stored at each grid point between timesteps) or as diagnostic variables (i.e. values are diagnosed from prognostic variables at each timestep). For example, a model with a 10 km horizontal grid resolution and a 5 minute timestep would require a prognostic representation of water vapour, cloud liquid water and ice/snow particles (ice with a fall speed of 1 m s^{-1} falls only 300m in 5 minutes), but rain could be represented as a diagnostic (rain with a fall speed of 10 m s^{-1} falls 3 km in 5 minutes).

(c) Examples of microphysical parametrizations

There exists an hierarchy of schemes used in climate, weather forecasting and cloud resolving numerical models of which some examples are described here. The simplest scheme has only water vapour as a prognostic and does not explicitly represent clouds but generates precipitation by removing any supersaturation in the atmosphere (Kessler, 1969). The Sundqvist (1978) scheme has only prognostic water vapour and treats cloud diagnostically by again assuming any supersaturation leads to condensation. In this scheme precipitation is modelled by instantaneously removing part of the condensate produced by a cloud scheme.

In the Smith (1990) and Rotstayn (1997) schemes, one prognostic variable is used to represent water vapour, cloud liquid water and cloud ice. The water vapour and condensate are combined to form a total water prognostic variable that is advected, and the vapour/condensate re-partitioning is performed diagnostically each timestep. The condensate is either liquid or ice or a mixture dependent on the temperature. Precipitation is treated diagnostically and is either rain or snow, dependent on temperature. The Smith scheme assumes that there is no ice supersaturation in the atmosphere in order to calculate the partitioning. This assumption is less valid than the assumption for water as significant ice supersaturations are commonly observed. Tiedtke (1993) describes a similar scheme in the ECMWF model which includes an additional prognostic equation for cloud fraction as well as condensate.

At the next level of complexity are parametrization schemes with one or more prognostic variables representing the ice phase, as well as water vapour and liquid water (e.g. Wilson and Ballard, 1999). Other schemes such as Fowler et al. (1996) split frozen water into two prognostics representing cloud ice and falling snow. More complicated parametrizations represent ice particle size spectra with two prognostic variables, the mass mixing ratio and number concentration (double moment schemes) (Ferrier 1994, Ikawa et al. 1991, Wang and Chang 1993, Ghan et al. 1997, Swann 1998, Seifert and Beheng 2001, Morrison et al. 2005). More complex bulk-microphysical parametrizations used in high resolution numerical models for convective scale studies include a further prognostic variable representing graupel and/or hail (Lin et al. 1983, Walko et al. 1995, Swann 1998).

3.2. Parametrization of Microphysical Processes

Returning to Eq. 1, the fourth term, representing the sources and sinks of a particle type, contains many microphysical processes, each requiring approximation and parametrization in a model. The number and complexity of the parametrized interactions will depend on the number of particle types represented in the model. However, the sources and sinks for the different particle types can be placed into three basic categories: phase nucleation, growth or evaporation by diffusion, and inter-particle collection. The equations that describe processes within each particle type are usually of a similar form. A brief overview of the main processes for cloud water droplets, rain and ice/snow particles follows:

(a) Nucleation

Nucleation of water from vapour, ice from vapour and ice from water are either homogeneous (i.e. a 'pure' system) or heterogeneous (i.e. a system containing impurities). Because of the high number of CCNs present in the atmosphere and the efficiency of water droplet nucleation at super-saturations less than 1%, it is generally assumed that all supersaturation condenses to form cloud droplets instantaneously. Nucleation of ice crystals can proceed via a homogeneous or heterogeneous process. Heterogeneous nucleation represents the growth of ice crystals on ice forming nuclei at any temperature below 0°C and is a strong function of temperature and supersaturation. In GCMs, it is often parametrized with simple threshold criteria dependent on temperature (Fletcher, 1962) and supersaturation (Meyers et al., 1992), unless the model has some representation of ice particle number concentrations and ice nuclei, in which case a more complex formulation can be used.

(b) Diffusion growth, evaporation and melting

Condensation growth (evaporation) of cloud water droplets will occur whenever the atmosphere is supersaturated (subsaturated) with respect to water. In terms of vapour density, the diffusional growth equation for an isolated particle at rest in a vapour field is

$$\frac{\partial m}{\partial t} = 2\pi D\chi(\rho_v - \rho_{vd}) \tag{2}$$

$\partial m/\partial t$ is the change of the particle mass with time, D is the particle diameter, χ is the molecular diffusion coefficient of water vapour in air, ρ_v is the ambient vapour density and ρ_{vd} is the vapour density at the particles surface. This equation is modified by kinetic effects, ventilation effects (for falling particles) and the effects of a non-uniform vapour field.

For ice particles, the terms deposition/sublimation describe the diffusion of vapour onto/away from the ice particle surface. An equation of the form of Eq. 2 applies to this process with the diameter D replaced by a capacitance term (with analogy to electrostatics) that takes account of the more complex geometry of ice crystals compared to spherical water droplets. The shape or habit (plate, column, dendrite) of the growing ice particle can be complex and depends on the temperature and vapour density excess over ice saturation in the growth environment. Ventilation effects, which act to increase the deposition/evaporation rate, also become significant if the particles are falling. One consequence of the fact that the saturation vapour density of ice is less than that of water is that ice crystals grow at the expense of any supercooled liquid water. This is often referred to as the Wegener-Bergeron-Findeison mechanism; if the atmosphere is below water saturation but above ice saturation, then liquid water droplets will evaporate and ice particles will undergo diffusional growth. Parametrization of the diffusion process usually follows Eq. 2 integrated over the particle mass spectrum with simple assumptions about particle characteristics.

The process of melting of ice/snow particles to liquid/rain drops is essentially a diffusion process governed by an equation of the form of Eq. 2, although of heat instead of moisture. This is a rapid process with a typical snowflake melting in a few minutes. It is for this reason that melting is often parametrized in models to occur spontaneously when the wet-bulb temperature is above 0°C, although it can be crucial to represent the melting process in marginal rain/snow situations when the temperature is just warmer than freezing. A significant proportion of the rain from mid-latitude frontal systems is due to the melting of ice/snow particles before they reach the ground.

(c) Collection processes

The governing equation for the collection process describes the volume swept out by the large particle and collection/coalescence of some proportion of the small particles in this volume, i.e.

$$\frac{\partial m_l}{\partial t} = E(\pi/4)D^2 n_s m_s (v_l - v_s) \qquad (3)$$

where m_l is the mass of the large collecting particle, E is the collection efficiency, n_s is the number concentration of small particles, m_s is the mass of the small particles and v_l and v_s are the fall speeds of the large particle and the small particles respectively. The efficiency term is a complex term that determines the proportion of particle collisions that result in a collection. However, for simplicity, a fixed collection efficiency is often used in microphysical parametrizations.

For a bulk-microphysical model, the above equation cannot be used directly as there is no explicit representation of the different particle sizes. Autoconversion describes the process of cloud water droplet collisions/coalescence to form raindrops; a process that is difficult to quantify and complicated by turbulence and other effects that are not fully understood. In a model parametrization, this process is often represented in a highly simplified form, as a linear function of cloud liquid water content with a threshold value before autoconversion can begin (Kessler 1969), or a non-linear function of liquid water (Sundqvist 1978, Seifert and Beheng 2001). A similar function is also sometimes used for conversion between ice categories in model parametrizations.

Aggregation (or autoconversion of ice) is the process of cloud ice particle collisions to form larger aggregate particles (snowflakes). Some form of Eq. 3 is often used for model parametrizations with more than one ice variable, otherwise the aggregation process is parametrized as a dependence of size spectrum on, for example, temperature.

Other collection terms include the accretion of droplets by raindrops as they fall through cloud and the collection of supercooled liquid water by ice particles (riming), both of which can be described by an equation of the form of Eq. 3, but is often simplified in bulk parametrization schemes with a single fall speed for the larger hydrometeor category and an assumption of zero fallspeed for the cloud category.

3.3. Discussion

This section has just touched the surface of microphysical parametrization. There is much more that could be said and there are other processes that are not mentioned above, such as particle break-up and ice particle splintering (Hallett and Mossop, 1974) that are often included in double-moment schemes. However, the purpose of this section was purely to give an outline of microphysical parametrization and to give a few examples of how the complex microphysical processes are simplified in atmospheric models.

To summarise, it is perhaps worth considering some issues that should be kept in mind when developing a microphysics parametrization:

(i) Accuracy vs. complexity vs. efficiency

Given a finite computing resource and uncertainties in our knowledge of microphysical processes, there is always a trade off between accuracy, complexity and computational efficiency of the parametrization. As an example, a computationally intensive, highly detailed scheme may be appropriate for an in depth high resolution study of a particular microphysical process or case study where the forcing is accurately specified, but not for a time-critical lower resolution operational NWP model with uncertainties in initial conditions and in the formulation of the dynamics and physics of the model. A parametrization with more degrees of freedom than can be constrained or understood and that is sensitive to uncertainties in the forcing can lead to a less accurate solution than a simpler parametrization with fewer degrees of freedom. Essentially, the complexity of the parametrization needs to be appropriate for the application. What “appropriate” means in this context is not always obvious and changes over time with our available computational resource and increasing knowledge of the microphysical system.

(ii) Traceability

Traceability is the completeness of information about every step in a process chain, or alternatively the ability to verify the history of a process by means of documented information. The representation of microphysics in a model is necessarily a simplification, an approximation of reality, so it is important to be able to trace these simplifications back to the source, whether this is theory, observations or more complex models that we believe to be closer to reality. This constrains the parametrization and gives confidence in the formulation and parameters.

(iii) Numerical formulation

The numerical implementation of the microphysics is important and particular care must be taken for models using relatively long timesteps (sequential vs. parallel, implicit vs. explicit). The microphysics should ideally be insensitive to model timestep as far as possible, or at least converge as the timestep is reduced. The numerics of hydrometeor sedimentation schemes have often been a source of significant sensitivity to timestep in models with long timesteps.

(iv) Quantifying uncertainty

There is uncertainty in every aspect of numerical modelling of the atmosphere, whether it is in the initial atmospheric state, uncertain parameters, uncertainties through approximations in formulation, or in numerical schemes. The evolution of a model can be sensitive to variations in the formulation within the level of uncertainty and it is therefore important to understand the uncertainties and the associated model sensitivity.

(v) Understanding impacts

Microphysics is just one aspect of a complex interaction of processes in the atmosphere (as discussed in the next section), and hence changing aspects of a cloud parametrization has consequences for the radiation and dynamics which may modify small-scale or large-scale circulations, which in turn feeds back on the forcing for the microphysics itself. It can be a challenge to understand the consequences of a specific parameter or formulation change in a GCM, but this is an essential part of the path to improving short-term forecasts and the long-term climate of the model. For example, Jakob (2002) describes the impact of changing ice fall speed in the ECMWF model; decreasing the fall speed of ice in the model increases the ice water path, which decreases the net radiative flux divergence (less radiative cooling) leading to a more stable atmosphere and a decrease in convective activity in the tropics.

4. Microphysics and Atmospheric Dynamics

4.1. Diabatic processes

As described earlier, microphysical parametrizations are often characterised by the number of water species that are represented and the number of prognostic variables that are used to describe each species (e.g. single-moment, double-moment, bin-microphysics). But in terms of the direct impact of a microphysics parametrization on the atmospheric state, it is the diabatic terms modifying atmospheric stability and hence the dynamical circulations that are important. An alternative view is to look at the scheme in terms of the diabatic processes of condensation/evaporation, deposition/sublimation and melting/freezing (radiative interactions could also be included but the focus here is on latent heating due to phase changes). The details of the parametrization scheme, such as the number of different classes of water species, determine the distribution of heating and cooling through the atmosphere. Figure 2 illustrates this concept. This way of thinking about the parametrization can help to relate simpler parametrization schemes to more complex multi-prognostic schemes and to understand the impact of a change to the parametrization. The most significant aspect of the choice of water species is the phase (water vapour, liquid, ice) and secondary to this is the split within a phase (cloud/rain; ice/snow/graupel) which acts to distribute the heating and cooling differently depending on fall speeds and particle characteristics, leading to different residence times in the atmosphere and differing rates of phase change (e.g. sublimation rate depends on particle shape).

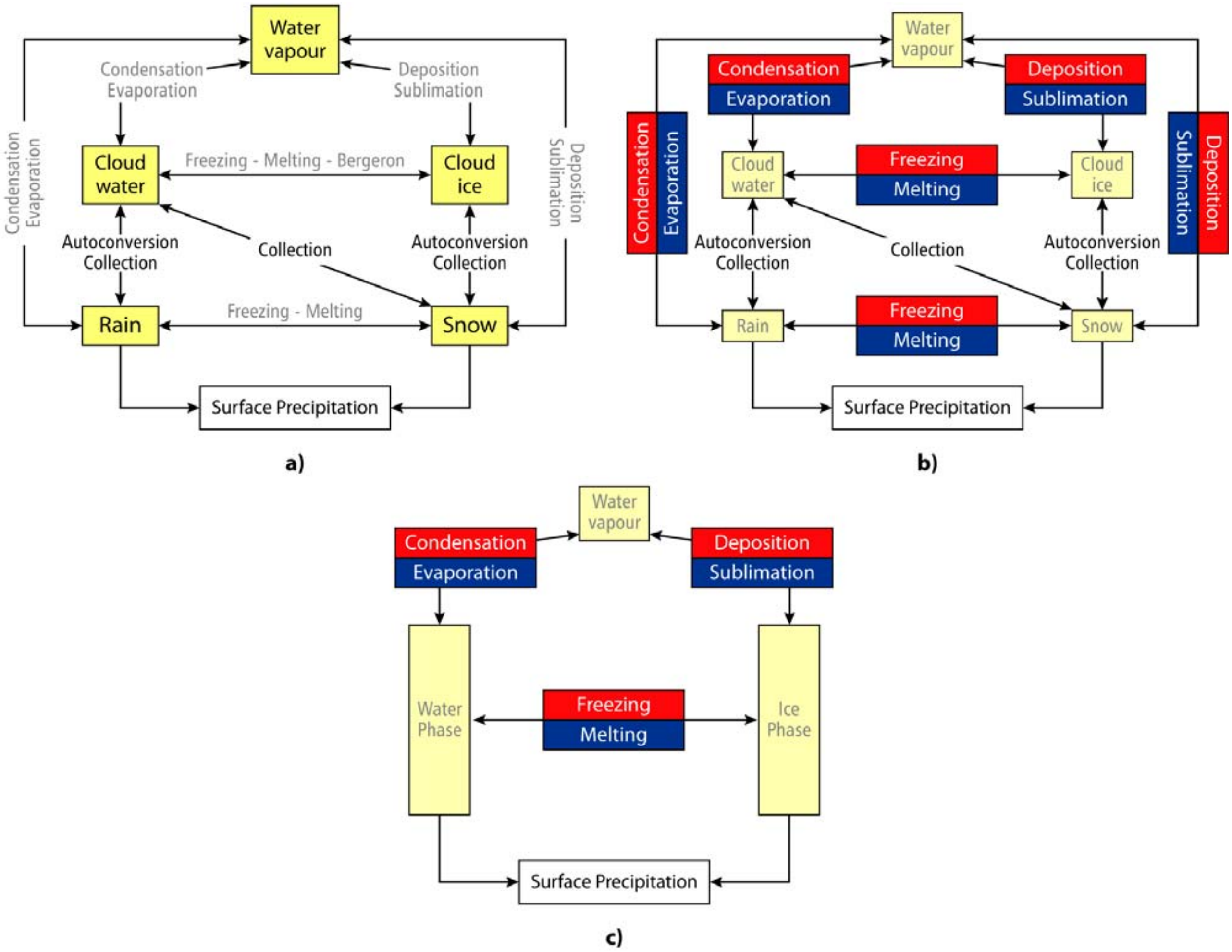


Figure 2: Schematic of a microphysical parametrization with water vapour and four categories of condensate/hydrometeors, (a) the "traditional" view, (b) highlighting diabatic processes, (c) a simplified view highlighting diabatic processes. Processes in red represent heating terms, processes in blue represent cooling terms and yellow boxes represent water categories.

The focus of this section is on the impact of microphysics on atmospheric dynamics through the heating and cooling due to phase changes. Particular emphasis is placed on cooling as this aspect is perhaps less well known than the impact of diabatic heating.

4.2. Microphysics and dynamics

The primary sources of diabatic heating in the atmosphere are from condensation of vapour to liquid water and deposition of vapour to ice and the importance of the effect of this heating on enhancing mesoscale and cyclone dynamics is well established (e.g. Robertson and Smith, 1983; Thorpe and Emanuel, 1985; Emanuel et al., 1987; Joly and Thorpe, 1989; Kuo et al., 1991; Mallet et al., 1999). However, there are also a number of observational and modelling studies of convective and stratiform clouds that suggest cooling due to the evaporation and melting of precipitation can also have a significant dynamical impact.

The evaporative cooling of precipitation beneath a sloping updraught can lead to a substantial cold pool and resulting destabilization of the atmosphere so that the cold air descends. In many situations these downdraughts have been associated with density current dynamics (Clarke, 1961; James and Browning, 1979; Carbone, 1982; Hobbs and Persson, 1982; Nielsen and Neilley, 1990). The downdraughts can flow away from the initial region of diabatic cooling, particularly when they reach the surface, and have a dynamical effect remote from the source region. For example, Moncrieff and Liu (1999) describe the initiation of convection by density currents away from the original source. Evaporatively induced downdraft outflow from thunderstorms can interact with the ambient shear and affect the life cycle of convection, longevity and dynamical structure of squall lines. Thorpe et al. (1982) suggest evaporation below the sloping updraught is a fundamental component of squall line development and maintenance. Tao et al. (1995) performed sensitivity experiments with a model of a mid-latitude squall line and found cooling by evaporation in the convective region was essential for maintaining a long-lived mid-latitude squall line cloud system. Johnson et al. (1993) performed a numerical modelling study of the role of ice in a convective storm. Including ice microphysics resulted in a longer lived storm because the ice with low fallspeed was advected by the upper level wind in the anvil away from the main updraught region, and the effects of evaporative cooling did not act to cut off the warm moist inflow. When the ice phase was turned off in the model, rain fell close to the updraught and cut off the inflow leading to a much shorter lifetime.

The latent heat of fusion at 0°C is roughly a factor of eight smaller than latent heat of ice sublimation so we might expect the dynamical effects of melting to be less significant than for ice evaporation. However, the cooling due to melting is confined to a layer just below the 0°C wet-bulb isotherm and may result in strong cooling with limited vertical extent. Also the cooling will occur whether the atmosphere is saturated or not, whereas evaporation is dependent on the atmosphere being sub-saturated. The particular spatial distribution of the cooling leads to particular phenomena associated with the melting layer. Findeison (1940) was the first to point out that cooling by melting snow can produce a 0°C isothermal layer that can be up to hundreds of metres deep and many authors have since described such a layer in observational and modelling studies. There is a significant dynamical effect of the isothermal layer due to the potential for reduced stability at the base of the layer and increased stability in the layer. Reduced stability near the base of the isothermal layer can lead to a super-adiabatic lapse rate and result in the triggering of convective cells (Findeison, 1940; Atlas, 1955; Atlas et al., 1969; Willis and Heymsfield, 1989). The increased stability in the layer means that stronger wind shears can be sustained in the layer providing a favourable level decoupling of flow in convection and frontal cyclones (Carbone, 1982; Stewart, 1984; Willis and Heymsfield, 1989; Stewart, 1990). Horizontal variations in the amount of cooling by melting snow in a baroclinic zone or at a rain/snow boundary can lead to thermal circulations (analogous to the sea-breeze effect) that could modify the precipitation (Lin and Stewart, 1986; Szeto et al., 1988a; Szeto et al., 1988b; Stewart and King, 1987; Stewart and Macpherson 1989). In a similar way to the effects of evaporative cooling there can be an impact on the development of squall lines and convective storms (Szeto and Cho 1994, Tao et al. 1995). For example, Tao et al. (1995) modelled the effect of melting processes on the development of mid-latitude squall lines. In the absence of melting processes, the simulated midlatitude squall system acquired the characteristics of unicell-type convection rather than the observed multi-cellular structure and affected the life-cycle of the system.

Although the processes of evaporation and melting both act to cool the atmosphere, there are differences between the characteristics of the two processes in terms of the magnitude of the latent heat change and spatial extent. The process of evaporation can occur at any height in the troposphere, but melting is confined to a region around the 0°C isotherm. Their different spatial characteristics can result in different dynamical effects and one process may dominate over another in different meteorological situations. For example, in mid-latitude mesoscale convective systems the cloud bases are high and the sub-cloud region is dry so the diabatic cooling effects of evaporation will dominate those from melting. In a hurricane, which generally has a lower cloud base and moist sub-cloud region, the effects of melting may dominate those of evaporation. In mid-latitude cyclones, different processes may dominate in different regions.

4.3. An Example: Microphysics and Frontal Dynamics

One example of the interaction between microphysics and dynamics is a mid-latitude frontal circulation. The first order impact of condensation heating and evaporative cooling in a cross-frontal circulation are illustrated with results from a simplified model using the semi-geostrophic equations. Secondly, a more realistic NWP model is used to highlight the impact of ice particle sublimation on a frontal circulation and sensitivity to aspects of the microphysics parametrization.

Figure 3(a) shows a schematic section of vertical velocity associated with the ageostrophic circulation across a front obtained using semi-geostrophic theory of frontogenesis (Hoskins and Bretherton, 1972) for a dry atmosphere. In this case, the potential temperature, θ, is conserved everywhere in the flow. For the dry case, the updraught and downdraught are symmetrical in the sense that they have the same magnitude and horizontal scale. We will refer to this as the "dry up dry down" case (DUDD).

Thorpe and Emanuel (1985) included diabatic heating (condensation and deposition) in the semi-geostrophic equations by assuming conservation of equivalent potential temperature, θ_e, in updraughts in an atmosphere of small stability to slantwise convection. They base this assumption on experimental field data that shows many frontal zones are nearly neutral to slantwise moist ascent. In the downdraughts, it is assumed that potential temperature, θ, is conserved so there is no representation of evaporative cooling; all precipitation is assumed to fall to the ground with no evaporation. This form of conditional parametrization can be referred to as "moist up, dry down" (MUDD). They performed a 2D numerical simulation of frontogenesis using the semi-geostrophic equations for this moist case and compared the MUDD case with the dry case. The effect of including diabatic heating in the updraught is to increase the rate of frontogenesis and collapse the updraught to a thin rapidly ascending sheet. Figure 3(a) and (b) illustrate the differences between the frontal circulations for the dry case (DUDD) and the conditional heating in the updraught (MUDD) case.

In order to represent the effects of evaporative cooling in a downdraught, a natural extension to the "moist up dry down" is the *unconditional* heating case, which makes the assumption of conservation of θ_e *everywhere* in the flow. This is referred to as the "moist up, moist down" (MUMD) system and results in symmetry between the updraught and the downdraught cross-front flows shown in Figure 3(c). The MUMD case is unrealistic as precipitation evaporation does not occur everywhere in the flow and the assumption of conservation of θ_e may not even apply in the evaporation zone. However, assuming that the approximation of a saturated downdraught is reasonable in regions of ice evaporation, a more realistic simulation could be obtained within this framework by a more appropriate parametrization of precipitation evaporation that limits evaporative cooling to the region beneath the frontal updraught.

Huang and Emanuel (1991) extend the work of Thorpe and Emanuel (1985) by including a more realistic representation of rain evaporation, but the impact of the sublimation of snow is likely to lead to stronger evaporative cooling than rain. Clough and Franks (1991) used a one dimensional model with microphysical parametrization of evaporation and fall speeds to show that for a given rate of precipitation, snow is much more effective at cooling the ambient air than rain. This is because ice has a much higher surface area for evaporation than rain for a given mass, and a much lower fall speed for a given rate of precipitation. This results in a higher rate of evaporation (and diabatic cooling) in a shallower depth. They also showed that in stable conditions evaporation of ice precipitation of 1-10 mm h^{-1} could maintain a near-saturated slantwise descending flow with a vertical velocity of 10-30 cm s^{-1}, typical of velocities found in frontal rainbands.

The role of snow sublimation on the frontogenesis process is investigated by Parker and Thorpe (1995) using the 2-dimensional semi-geostrophic model and a simple parametrization of sublimation cooling. The parametrization of diabatic heating is the same as that used by Thorpe and Emanuel (1985) based on the assumption of conservation of θ_e and small stability to slantwise convection in the updraught. The sublimation cooling parametrization also uses this assumption in the downdraught (as in the MUMD case) but the cooling is limited to a prescribed region beneath the frontal updraught to represent the region of snow sublimation more realistically. Instead of a symmetric solution, a narrow downdraught forms with a maximum vertical velocity similar in magnitude to the updraught. Figure 3(d) illustrates the impact on the vertical velocity in comparison with the MUMD and MUDD cases.

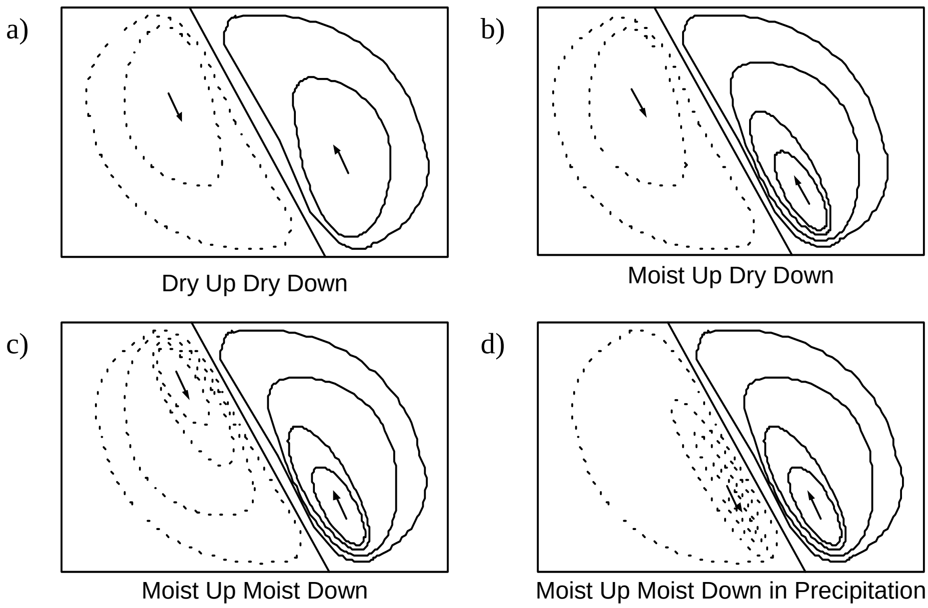


Figure 3: Schematic cross-front section of vertical velocity updraughts (solid lines) and downdraughts (dotted lines) associated with frontogenesis predicted by the (a) dry semi-geostrophic (s-g) equations (DUDD),(b) s-g equations with moist updraught and dry downdraught (MUDD), (c) s-g equations with moist updraught and moist downdraught (MUMD) (Thorpe and Emanuel, 1985), (d) s-g equations with moist updraught and moist downdraught limited to a region beneath the frontal updraught to represent the effects of precipitation evaporation (Parker and Thorpe, 1995). The x-axis is horizontal distance across the front (0-1000 km) and the y-axis is height from the surface to the tropopause (0-10 km).

This idealised example illustrates the role of condensation (deposition) heating and evaporation (sublimation) cooling on the dynamics of a mid-latitude front. The characteristics of the updraught and downdraught are significantly modified becoming stronger and narrower. Most of the source of the diabatic cooling in the downdraught is at temperatures below freezing in mid-latitude fronts and hence is dominated by sublimating snow (Forbes and Hogan, 2004). The fact that the magnitude and depth of the cooling is dependent on the sublimation rate and fall speed of the snow suggests the dynamical impact will depend on details of the model microphysical parametrization. There may also be microphysical-dynamical feedbacks that are not represented in the idealised simulations described above and the semi-geostrophic equations break down when the moist potential vorticity is small and so such a balanced system is less applicable for studying systems with large diabatic forcing.

Forbes (2002) and Forbes and Clark (2003) investigate the role of snow sublimation in frontal dynamics using the UK Met Office Unified Model at 12 km grid resolution, a primitive-equation model which includes a prognostic representation of cloud liquid water and ice/snow with advection and sedimentation, and a cloud parametrization representing the key microphysical process rates (Wilson and Ballard, 1999). A simulation of an idealised front in a 3-dimensional domain is generated from an initial state consisting of a thermal gradient in the horizontal and an upper level jet in thermal wind balance, moist neutral stability and a deformation wind field that is forced at the lateral boundaries of the domain. Figure 4 shows a vertical cross-section across the front after a few hours of simulation, comparable to the schematic cross-sections in Figure 3. Figure 4(a) shows the region of frozen cloud created by the sloping frontal updraught. The frontal updraught and the downdraught beneath the frontal surface are shown in Figure 4(b). The narrow intense downdraught is particularly noticeable. Figure 4(c) shows the latent heating rate cross-section with heating from ice deposition in the updraught and cooling below from snow sublimation both acting to enhance the cross-frontal circulation. A sensitivity experiment with the cooling due to snow sublimation artificially set to zero results in the downdraught weakening considerably which also feeds back to weaken the updraught at lower levels. This highlights the sensitivity of the downdraught to the diabatic cooling, which will depend on the sublimation rate and terminal fall velocity of snow parametrized in the model. This in turn depends on assumptions of particle size distributions, particle densities and particle morphology (through the capacitance term in the sublimation rate equation). Additional sensitivity experiments varying the sublimation rate and terminal fall speeds by a factor of two (arguably within the level of uncertainty) results in significant changes in the strength and depth of the downdraught (not shown here).

Forbes and Clark (2003) perform the same type of sensitivity experiments in a real case study of a mid-latitude cyclone with associated fronts using the same version of the UK Met Office Unified Model with a 12 km grid resolution. Figure 5 shows the area of the low pressure centre over the North Atlantic to the west of Ireland with a warm front extending to the east and a cold front extending to the south. The vertical velocity is shaded and shows the ascent associated with the cyclone centre, the warm and cold fronts and the band of descent to the west of the main cold front. The two panels in Figure 5 show the difference when the fall speed of snow is halved and doubled from the reference value respectively. Reducing the fall speed leads to increased ice/snow water contents, significantly increasing the sublimation cooling in the frontal downdraught, strengthening the downdraught and acting as a positive feedback on the main frontal updraught and a secondary front further to the west (increasing ice amounts further). However, note that the impacts are confined to the frontal scale and do not significantly affect the cyclone scale.

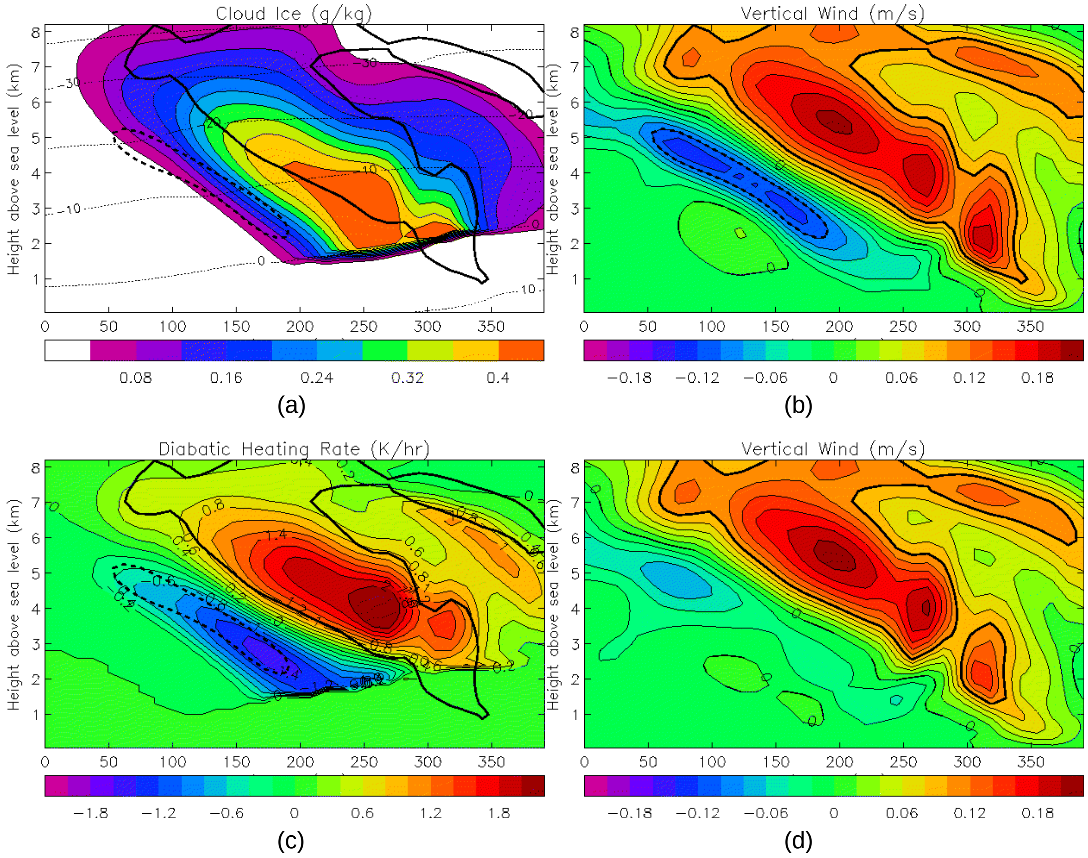


Figure 4: Cross-section through an idealised front using the UK Met Office Unified Model at 12km resolution; Reference simulation (a) cloud ice (shading) and temperature (thin contours),), (b) vertical wind velocity (also thick contours in a,c) (c) latent heating rate (shading). (d) Vertical wind velocity from the simulation without the cooling due to snow sublimation. The horizontal axis is distance in km.

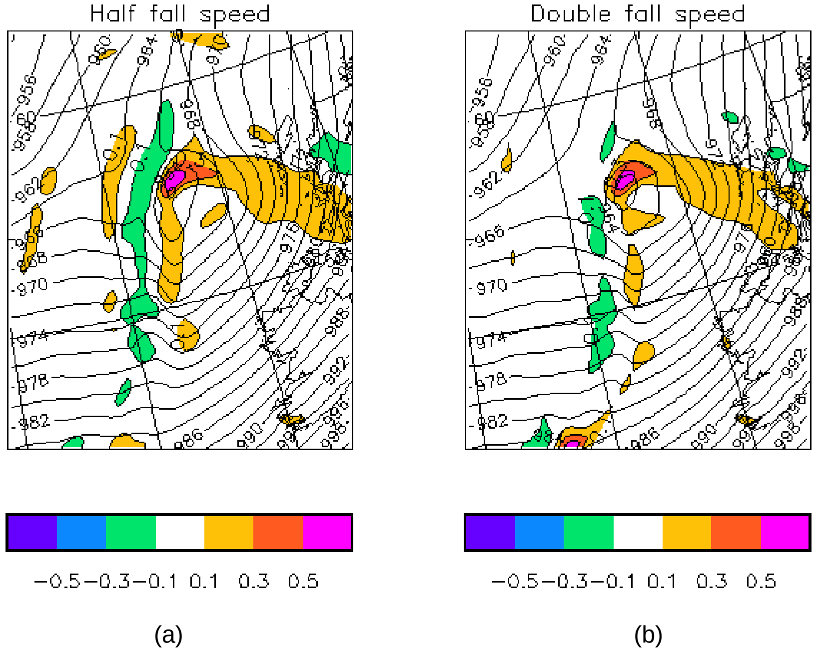


Figure 5: Plan view of vertical velocity (shading - m/s) at 850hPa and mean sea level pressure (contours, hPa) from Met Office Unified Model 12km resolution simulations for a low pressure case study to the west of Ireland with (a) half and (b) double the reference ice terminal fall speed.

The purpose of this section has been to show an example of the interactions of microphysics and dynamics through diabatic heating and cooling and to highlight the sensitivity of the dynamics, in the context of an operational meso-scale model forecast, to the details of the microphysics scheme parametrization; details which may have significant uncertainty associated with them.

5. Discussion and Conclusion

This paper has discussed the parametrization of cloud and precipitation microphysical processes in atmospheric numerical models with a focus on the interaction of microphysics with atmospheric dynamics through latent heat processes. The atmospheric system, and our representation of this system in numerical models, is a complex web of interactions and feedbacks between many different processes and microphysics is just one component. An example of the role of evaporative cooling in mid-latitude fronts highlights the way in which details of the microphysical parametrization in a NWP model can affect the dynamics and subsequent forecast skill of the model. There are many other examples relevant for NWP and climate models, from the small-scale impacts on the development of individual tropical convective cells to the large-scale impacts of diabatic heating in mid-latitude storm-tracks. The main point is that we should try and understand changes in the microphysics, not just in terms of cloud cover, water contents and particle size distributions leading to radiative and hydrological impacts, but also in terms of dynamical impacts through changes in diabatic processes. For parametrization development we should aim to:

- quantify uncertainty and model sensitivity to this uncertainty in order to understand the impacts of microphysics on the model system as a whole (dynamical, radiative, hydrological) and know where to target effort to further constrain the parametrization.
- ensure that the parametrization has traceability to observations, theory and to more complex models to appropriately constrain and give confidence in the formulation and parameters of the parametrization.
- strike the appropriate balance between the required level of detail of the parametrization, appropriate complexity of the formulation and computational efficiency, depending on the degree of understanding of the processes and the purpose and application of the model.
- ensure the numerical implementation of the microphysical parametrization is robust as the timestep of the model is reduced.

There are many areas of current research and prospects for further development of microphysical parametrization. Regarding microphysical issues, there is potential for improving our understanding of almost all microphysical phenomena through laboratory studies, field observations, theory and detailed modelling, from warm-rain production to ice crystal morphology. However, a very active field of research at present is the interaction of aerosols and microphysics, particularly ice nucleation processes and the direct and indirect aerosol effects on radiation, with interest in the latter from the perspective of climate change. Further research is needed for processes in super-cooled and mixed-phase cloud. For example, thin layers of supercooled-liquid water are common (Hogan et al., 2004), can often be radiatively important and yet current models are generally unable to represent this type of cloud.

In many clouds, it is the small-scale dynamics that provides a significant part of the forcing for the microphysics through generation of supersaturation and subsaturation, yet these small-scale motions are not represented, or are represented in a crude way, in large-scale models. The parametrization of sub-grid heterogeneity (whether humidity, condensate, temperature or vertical velocity) is an important concept in NWP and climate models which has not been discussed here (as it is the topic of Tompkins' paper in this volume) but can be a significant source of model error. Although there has been progress in building alternative sub-grid cloud schemes (e.g. Tompkins 2002) further research in this area is needed.

Unifying cloud microphysical assumptions across model parametrizations (stratiform cloud / convection / radiation) should be a goal we should be working towards, not just for consistency, but also for representing interactions and feedbacks in a more realistic way. For example, many radiation parametrization schemes use a fixed effective radius or cloud condensation nuclei concentration and there is scope for making a direct link with information from the cloud scheme, provided that the latter is able to provide this information with sufficient accuracy.

Data assimilation of remote sensing observations continues to be an active area of research with the increasing number and variety of space-bourne and surface remote sensing instruments. The approach of forward modelling to emulate the observed quantities relating to cloud (cloud-affected radiances, radar reflectivity, lidar backscatter) could benefit from improvements in the representation of microphysics in the model, for example particle size distributions, if this can be predicted with some skill.

The advent of active radar and lidar instruments on board satellites, such as Cloudsat and CALIPSO, brings with it a wealth of information on the vertical distribution of cloud properties around the globe (see Stephens paper in this volume). Ground based remote sensing sites are also providing long time-series of cloud profile retrievals (ARM, CloudNet) and these new data sources can be used to extract information on microphysical properties and even infer information about microphysical processes. Validation is an integral part of parametrization development and extracting as much as we can from these new instruments as well as from the passive satellite instruments and in situ aircraft observations will continue to be an active and valuable area of research.

It is a challenge to understand and simplify a complex system of interactions that spans a range of scales from microns to kilometres, over nine orders of magnitude, but even simple parametrizations have a degree of success in describing the first order impact of microphysical processes on atmospheric dynamics. Increased understanding through a combination of theory, observations and detailed numerical modelling has and will continue to lead to further refinement of microphysical parametrization schemes for the ongoing improvement of numerical weather prediction and climate models.

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