The value of hierarchies and simple models in atmospheric research

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Key Points:

- Simple models have advanced our understanding of the atmosphere. Key benchmark models are identified.
- Hierarchies help address open research questions. We focus on how they have improved understanding in circulation, clouds, and convection.
- Model hierarchies are commonly referred to but remain poorly defined. We identify three principles to order models in the atmospheric sciences.

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Abstract
Models – both simple and complex – have enabled our understanding of the atmosphere. In this review, we highlight the complementary relationship between simple and complex models in addressing key questions in atmospheric science. The systematic representation of models in steps, or hierarchies, link our understanding from idealized systems through to the state-of-the-art models, and ultimately our atmosphere. Three interconnected principles characterize model hierarchies of the atmosphere. *Dynamical hierarchies* allow us to isolate and explore the importance of temporal and spatial scales on the governing equations. *Process hierarchies* allow for the segregation and systematic integration of important atmospheric processes. Finally, the *hierarchies of scale* allow for the systematic exploration of dynamical and physical processes via physical domain choices and numerical resolution.

We center our discussion on the circulation of the atmosphere as well as its interaction with clouds and convection, focusing on areas where simple models have had a significant impact. Our confidence in climate model projections of the future is embedded in our efforts to ground the climate predictions in fundamental physical understanding. This understanding is, in part, possible due to the hierarchies of idealized models that afford the simplicity required for understanding complex systems.

1 Introduction

All models are wrong but some are useful [*Box*, 1978]. The statistician George Box succinctly made two important points at a workshop on statistical robustness held four decades ago. First is the reminder that most models, even our most sophisticated, are far from reality. With respect to the atmosphere, we have little hope of achieving a model capable of explicitly simulating all processes from the global to microphysical scale, at least in the foreseeable future. It is his second point, however, that is most germane to this study: we can learn, understand, and make predictions with *some* models. The aim of this review is to identify some of the deliberately wrong (or, put differently, idealized) models that have proven useful for understanding and predicting the behaviour of our atmosphere, and their organization into hierarchies that connect them with our most complex modeling efforts.

The notion of a 'hierarchy in climate models' is by no means new, with perhaps the first explicit discussion to be found in *Schneider and Dickinson* [1974]. They discussed the hierarchy of models available then and commented that 'solid progress in understanding … climate change will require steady development of an almost continuous spectrum or hierarchy of models of increasing physical or mathematical complexity'. This sage advice was evidently not well heeded, for a decade later *Hoskins* [1983] noted the 'unhealthy' trend toward building models which are disconnected from one another and the real world, advocating, like Schneider and Dickinson, for a spectrum of connected models to provide a complete and balanced approach. Two and then three decades on from that, *Held* [2005] and *Held* [2014] highlighted the widening gap between our understanding of the atmospheric circulation and the increasing complexity of global circulation models. He argued for the study of *elegant* models that are simple enough to answer our key scientific questions. Relatedly, *Nof* [2008] criticized the trend in climate modeling for higher resolution over increased understanding, and the danger of regarding comprehensive models as 'truth'. Or, as argued by *Polvani et al.* [2017], 'Earth system models may be good for simulating the climate system but may not be as valuable for understanding it'.

The lack of reproducibility in modeling experiments also suggests a need for models of varying complexity. *De Verdiere* [2009] suggested that whereas modeling Intercomparison Projects (MIPs) are useful for identifying variability in a process's response to forcing, we need to dig deeper to diagnose and understand this response. A potential alternative to the MIP approach is, then, to use a hierarchy of climate models to gain physical understanding, as opposed to (or perhaps in addition to) trying to converge to observations. *Jeevanjee et al.* [2017] further emphasize that model hierarchies motivate hypothesis testing, specifically by allowing the formulation of mechanism denial studies.

Given this seemingly almost universal agreement on the need for model hierarchies in the atmospheric sciences one may wonder why they are not in more widespread use. Part of the answer is that comprehensive (or 'high-end') models have been, in spite of the criticism sometimes leveled at them, enormously successful in many respects – weather forecasting being the most obvious, but not the only, example. In many ways they have outstripped our theoretical understanding, and the need to have simpler models (or indeed for any form of understanding) for a good simulation is not always apparent [as discussed in *Vallis*, 2016]. Two other issues also provide impediments to hierarchy building. The first is that identifying and agreeing upon the models that are most appropriate and useful is not trivial, and a unique or 'best' hierarchy is neither possible nor needed. Whereas there is some agreement at the very complex end of the hierarchy (models with as complete a set of physical processes as possible) and at the very simple end (e.g., very simple energy balance models), there are many paths between them, some more sensible than others. As emphasized by *Held* [2005], in biology there are natural intermediate systems (e.g., the fruit fly, the mouse) that can be used to understand the human body, there is no analogue of that in the atmospheric sciences, at least as regards Earth's atmosphere. The second impediment is the practical issue of *meaningfully* sharing models across research groups, which is helpful for establishing a common hierarchy. It is one thing to provide computer codes, but quite another to allow others to effectively use and adapt it for new research purposes.

In this review, we focus on the first issue, identifying a number – though by no means not all – of the models that form a hierarchy in atmospheric research. (A perspective on the related but somewhat broader subject of climate hierarchies is to be found in *Ghil and Robertson* [2000].) We begin in Section 2 by identifying three principles to characterize model hierarchies: dynamics, processes, and scales. We then briefly discuss the dynamical hierarchies in atmospheric fluid flow in Section 3. In Section 4, we explore a process hierarchy of general circulation models, where the diabatic processes driving the thermodynamic equation are systematically advanced. We term this sequence of models a 'diabatic hierarchy'. We then focus on the models that helped us understand the circulation of the midlatitudes, middle atmosphere, and tropics in Sections 5, 6, and 7, respectively. Finally, we focus on the unresolved processes at the forefront of atmosphere research, tropical convection in Section 8 and clouds in Section 9. After synthesizing the key results of the paper in Section 10 we conclude our review in Section 11. Our focus is on *dynamical* models and we do not discuss such things as energy balance models, important as they are. Nor do we discuss coupled atmosphere-ocean models, and thus (among other omissions) we do not discuss the Cane–Zebiak El Niño model, one of the most influential simple models in all of climate science. Elsewhere our choices are, given the limited space, perhaps a little arbitrary, with (for example) only a brief mention of the quasi-biennial oscillation.

Although we shall not discuss it further, there has been progress in the second potential impediment. The Portable University Model of the Atmosphere (PUMA), introduced by *Fraedrich et al.* [2005], was a pioneering effort to make atmospheric circulation models more configurable and user-friendly. New software and resources are now becoming available to enable more systematic use of atmospheric models developed by several of the major modeling centers. A suite of models based on codes developed by the Geophysical Fluid Dynamics Laboratory (GFDL) was extended by *Vallis et al.* [2018a] to form the open-source Isca framework, which includes a modern user interface that allows a wide range of parameterizations and configurations, and that can be run on Macs, PCs, Linux boxes and supercomputers. And as part of the next model release, the Community Earth System Model (CESM) will include two simple models: an aquaplanet and dry dynamics core [*Polvani et al.*, 2017]. Both configurations are set up and run with the same set of commands as the fully-coupled earth system models. An atmospheric single-column model capability is also being added, and a procedure for incorporating simplified atmospheric physics packages has been developed to allow further configurations to be added in the future.

2 Principles guiding model hierarchies

What does it mean to be a 'simple' or 'idealized' model, and how does such a model sit within a 'hierarchy'? The first and perhaps most important single point to make is that a useful hierarchy involves a *connected* sequence of models; aside from the end members, each model should ideally be connected to models of greater or lesser complexity. Simplicity is then defined relative to other members of that hierarchy. For the purposes of this review, we focus on models that are *deliberately* simplified, and a conscious effort to limit model complexity is a first step towards establishing models in a hierarchy. Categorizing models in terms of their complexity, as expressed by *Bony et al.* [2013] in Figure 1a, is one useful way for describing the configuration options and assessing which model is appropriate in which context. Relatedly, *Jeevanjee et al.* [2017] describes the climate model hierarchy, see Figure 1b, as a Cartesian product space of individual hierarchies that can be grouped into dynamics (fluid and rotation processes), boundary layer forcing (ocean and surface processes) and bulk forcing (e.g. convection and radiation). We propose an alternative, but complementary, description, based on the idea that there are three principles which help organize the formation of model hierarchies within the atmospheric community. These three principles are dynamics, processes, and scales, as illustrated schematically in Figure 2 and discussed further in Sections 3 and 4.

Dynamical hierarchies allow us to isolate and explore the importance of different temporal and spatial scales on the governing equations. As detailed in Section 3, hierarchies of dynamical equations were instrumental developing effective numerical models.

Process hierarchies allow for the stepwise integration of important atmospheric processes into the governing equations of the fluid flow. Processes are sometimes integrated directly from first principles, as with radiative transfer, but often must be parameterized, as with cloud microphysics. A cloud microphysics example of the process hierarchy is the progression from pure radiative equilibrium to radiative-convective equilibrium (RCE) in the absence of clouds, and then a full cloud resolving model (CRM).

Implicit in both dynamical and process hierarchies are *hierarchies of scale*, where the choice of physical domain and numerical resolution allows for the systematic exploration of different dynamical and physical processes. Here, there are practical trade offs between scale and complexity due to the computation expense. For example, consider the contrast between high resolution simulations of a limited domain model that seeks to accurately model convection compared to a moderately resolved global model that seeks to estimate the subgrid-scale convective processes.

Almost all theory and modeling efforts can be classified into a hierarchy of some form, so attempting to catalogue *all* the hierarchies is a hopeless task. In the remainder of this paper, we selectively highlight examples of model hierarchies, specifically those that include simple models and that have advanced our understanding of specific aspects of the atmosphere's behavior. We focus on these models not because they necessarily optimally cover the complexity spectrum, but rather because there is a significant body of literature on them, establishing their impact. We make no effort to discuss comprehensive GCMs in detail, but they should be seen as an end-member in the hierarchy of atmospheric models; that is, they are *part* of the hierarchy, not separate from it.

3 The Dynamical Equation Hierarchies and Hierarchies of Scale

The governing equations of atmospheric fluid flow offer a textbook example of a model hierarchy, and we refer the reader to (for example) *Gill* [1982] or *Vallis* [2017] for more comprehensive discussion. The atmosphere is most accurately described by the rotating Navier-Stokes equations that are capable of describing a Beethoven symphony or the flow about the wing on a supersonic jet as well as the slow dynamics of climate. Although comprehensive models used for quantitative prediction may use the full equations of motion, rarely does the need arise to describe all the scales simultaneously and the very fast dynamics are an unneeded complication for describing the motions relevant for climate scales. Thus, more idealized models simplify these equations in one way or another to filter out those unwanted modes. The simplifications follows two principles, the first being focused on the scales of interest, the second on the treatment of compressibility. Regarding the first, a deliberate focus on motions of certain spatial and time scales allows one to filter out less important (and generally faster) motions; indeed the first successful numerical weather forecast by *Charney et al.* [1950] was done with the two dimensional (latitude-longitude) *quasi-geostrophic* (QG) vorticity equations. The quasi-geostrophic equations are a filtered set of governing equations appropriate for synoptic scale motions in the presence of strong rotation and stratification [*Charney*, 1948]. The key advantage of the quasi-geostrophic equations is to filter out the presence of not only sound waves, but also gravity (or buoyancy) waves, leaving only the slowest Rossby modes of the system, which dominate day-to-day weather in the midlatitudes.

The strong rotation assumption of the QG equations, however, is not appropriate in the tropics. In addition, one must assume *a priori* the stratification of the atmosphere, limiting important feedbacks between the circulation and stratification. The so-called *primitive equations* offer an equation set appropriate for the whole globe (at least if the atmosphere is appropriately shallow) and allow the dynamics (and thermodynamics) to influence the stratification. The primitive equations assume that the atmosphere is in hydrostatic balance, which is justified when the vertical scales of motions are much less than horizontal length scales as found in our 'thin shell' atmosphere. The primitive equations permit gravity wave motions (although they are generally under-resolved at the resolution of most climate models), but hydrostatic balance filters out vertically propagating sound waves, allowing efficient numerical representation of the circulation. Consequently, the primitive equations are still used in most climate models. The aspect ratio of atmospheric convection violates the assumptions required for hydrostatic balance, requiring the use of non-hydrostatic equations in cloud resolving models. Non-hydrostatic effects, however, do not become generally important until a model can resolve convection – involving scales on the order of kilometres – which are currently computationally out of reach for the purposes of routine climate prediction.

Scale consideration also influence the choice of geometry. The use of a locally Cartesian grid, explicitly neglecting spheric effects, can be justified when motions of interest are sufficiently small relative to the radius of the earth. Use of a Cartesian geometry is generally associated with an idealized treatment of rotation, e.g., the use of a f or β-plane, where the effective rotation in the vertical plane is assumed to be constant, or vary linearly, within the domain.

A second principle of the equation hierarchy focuses on the relationship between temperature, pressure, and density, simplifying impacts of compressibility and thermodynamic processes. Barotropic dynamics (i.e., 2-dimensional horizontal, where the 'barotropic' here refers to motion independent of pressure or height) eliminate the impact of density fluctuations within the flow, and so focus exclusively of the role of waves and vorticity, as discussed in Section 5.1. The shallow water equations permit the simplest inclusion of density effects, which are modeled by the thickness of the fluid layer. Multilayer shallow water models begin to capture the influence of temperature on density.

The *Boussinesq equations* provide an idealized framework to capture the impact of temperature on density, but avoid the impact of compressibility on the density, and can be used in the atmospheric flows over limited vertical height. They can be applied with or without hydrostatic balance, depending on the scales of interest. The *anelastic equations* keep the spirit of the Boussinesq approximation and eliminate sound waves, but capture the compressible effects that are important on larger vertical scales. They are the equations of choice for some cloud resolving models, such as the cloud resolving System for Atmospheric Modeling [SAM; *Khairoutdinov and Randall*, 2003].

4 The Diabatic Hierarchy: Global Models with Varying Physics and Boundary Conditions

Our third principle focuses on atmospheric processes. The atmosphere is set into motion by the uneven heating of the planet by the sun, both vertically, as most solar radiation is absorbed at the surface, and horizontally, as the tropics receive more energy than the poles. There are a number of ways to formulate the thermodynamic equation of the atmosphere, for instance,

$$c_p \frac{D\theta}{Dt} = \frac{\theta}{T}\dot{Q} \tag{1}$$

where c_p is the specific heat capacity of air, $D\theta/Dt$ the material (total) derivative of potential temperature θ, and T the temperature. The seemingly innocuous $\dot{Q}$, the net heating rate, per unit mass, hides all the complex 'atmospheric physics', or non-conservative diabatic processes that drive the circulation. These diabatic processes are a central challenge in modeling the atmosphere. Many processes occur on scales much smaller than the typical grid resolutions of global models and so must be represented by sub-grid parameterizations. The choices made in their design lead to a variety of parameterizations, and hence models. Below we attempt to rationalize these choices in terms of a hierarchy of diabatic processes, which we further separate into two components. First we focus on the thermodynamics of the atmosphere in Sections 4.1-4.3, and then the choice of boundary conditions in Section 4.4.

4.1 Dry General Circulation Models

We may consider the base of the diabatic hierarchy to be collection of 'dry' GCMs. These models crudely approximate all diabatic processes by a simple Newtonian relaxation of temperature (i.e. Newtonian cooling) to an equilibrium temperature profile, T_{eq} so that

$$\dot{Q} = -c_p \frac{T - T_{eq}}{\tau}, \tag{2}$$

where τ is the relaxation time scale for the equilibrium state. One of the first models with Newtonian relaxation and Rayleigh friction, needed for surface momentum exchange, was *Hoskins and Simmons* [1975], but now a broad range of modeling groups use similar formulations.

Held and Suarez [1994] proposed a structure of T_{eq} and τ that produced a quite realistic climatology. Briefly, the troposphere is relaxed towards a state approximating a radiative-convective equilibrium, with near moist-neutral stratification in the vertical, but strong meridional temperature gradients. Above the tropopause, the atmosphere is simply relaxed towards an isothermal state. Another possible forcing is pure-radiative equilibrium, for example as described in [*Schneider and Walker*, 2006] and which employs an equilibrium temperature profile that is explicitly unstable, and hence more representative of the atmosphere. This forcing requires the additional use of an idealized convection scheme to explicitly mimic the stabilizing effect of latent heating by moist convection.

Dry GCMs have been widely used due to their balance between simplicity and realism. Because the static stability and tropopause height are internally determined, these models can produce richer dynamical behavior than the QG model. They have been used to study, among other things, the tropopause height [*Schneider*, 2004; *Zurita-Gotor and Vallis*, 2011], the relationship between tropospheric depth and jet latitude [*Lorenz and DeWeaver*, 2007], the storm tracks [*Chang*, 2006], the natural variability of the midlatitude circulation [*Gerber and Vallis*, 2007, 2009], fluctuation-dissipation theory [*Ring and Plumb*, 2007, 2008], the response of the atmosphere to global warming [*Butler et al.*, 2010], and the sensitivity of baroclinic eddies to the vertical structure of baroclinicity [*Yuval and Kaspi*, 2016]. Dry GCMs have also provided insights into the degree to which the Hadley circulation strength and extent are affected by extratropical eddies [*Kim and Lee*, 2001; *Walker and Schneider*, 2006; *Korty and Schneider*, 2008; *Sobel and Schneider*, 2009] and differ from classical angular momentum conserving theories [*Held and Hou*, 1980].

Dry models of the tropical atmosphere have also proven surprisingly successful in simulating phenomena, such as monsoons [*Schneider and Bordoni*, 2008], equatorial waves [*Potter et al.*, 2014] and tropical cyclones [*Mrowiec et al.*, 2011], where latent heating might have been thought to be a necessary ingredient in a minimal model. Dry GCMs with passive water vapor—subject to advection and condensation—have also proven useful for studying the climatological humidity distribution [*Galewsky et al.*, 2005; *Ming et al.*, 2017] in isolation from the active effect of latent heat release. These models have also been used to explore the impact of tropospheric circulation to changes in stratospheric water vapor and tropospheric warming [*Tandon et al.*, 2011, 2013], and the impact of cloud-radiative forcing on the extratropical jet stream [*Voigt and Shaw*, 2016].

4.2 Idealized moist general circulation models

A starting point for the inclusion of latent heating in global models is to interactively simulate the atmospheric hydrological cycle in a so-called idealized 'moist GCM'. Interactive here means that water vapor is prognostically evolved as an active tracer that is subject to advection by the atmospheric circulation, surface evaporative sources, and atmospheric condensation sinks. A widely used *benchmark* model along those lines is that of *Frierson et al.* [2006] in which the diabatic forcing is greatly idealized, in three main respects. First, convection is parameterized using the simple Betts–Miller scheme that relaxes a convectively unstable vertical profile back to a moist adiabat (i.e. a stable profile) in a manner similar to (2). Second, there are no cloud processes. Third, the model uses a so-called 'gray radiation' scheme, in which the spectrum of infra-red radiation is treated as a single band, with a single optical thickness, independent of the prognostic water vapor. As a result, the model of *Frierson et al.* [2006] does not include water vapor radiative feedbacks, but oes capture its influence on latent heat transport.

Recent efforts have begun to add water-vapor radiative effects without explicitly including cloud microphysics. This can be accomplished by replacing the fixed optical depth gray-radiation scheme with a one that depends water vapor content, and including, or not, cloud radiative feedbacks. Simple radiation schemes along these lines include those of *Byrne and O'Gorman* [2013] and *Geen et al.* [2016], which are gray and two-band in the infra-red, respectively. A more complete radiative scheme is included in *Jucker and Gerber* [2017] who study the tropical tropopause layer (still without clouds) and in *Merlis et al.* [2013], who study the impact of precession on climate with a prescribed cloud distribution.

Idealized moist GCMs with interactive hydrological cycle [*Frierson et al.*, 2006] have been used extensively. They have been used for many aspects of tropical climate change including the Hadley cell extent and strength [*Schneider et al.*, 2010], the Walker circulation weakening with warming [*Merlis and Schneider*, 2011; *Wills et al.*, 2017], a minimal model of monsoon transitions [*Bordoni and Schneider*, 2008], and as a framework for understanding the response of the Intertropical Convergence Zone (ITCZ) to extratropical and tropical thermal forcing [*Kang et al.*, 2009; *Bischoff and Schneider*, 2014].

4.3 Comprehensive Atmospheric General Circulation Models

Atmospheric General Circulation Models (hereafter AGCMs), which serve as the atmospheric components of Earth system models, include state-of-the-art representations of all the diabatic processes. Idealized AGCMs and comprehensive AGCMs generally share the same dynamical core but the treatment of the diabatic processes varies. Comprehensive AGCMs include more realistic representations of diabatic processes, but can still be stratified within a hierarchy in terms of the number and sophistication of processes they represent (the treatment of aerosols, chemistry, convection, surface friction, radiation, clouds), and through the choice of boundary conditions, as discussed in the following subsection.

A common difference between the idealized GCMs and comprehensive models is the treatment of clouds, as discussed in Section 9. Interactive clouds greatly increase the complexity of an atmospheric model as both macrophysics (cloud fraction, cloud optical properties, etc.) and microphysics (broad transitions of water phases) may be included in a full model. Cloud microphysics are strongly influenced by atmospheric aerosols, requiring at least a minimal treatment of these processes as well. Many modern comprehensive GCMs include additional prognostic tracers that allow for interactive atmospheric chemistry and aerosol-cloud effects. Certainly, there is scope for deliberate simplification of these processes, emphasizing that that model hierarchies naturally evolve in time as the comprehensive end-members evolve: a state-of-the-art AGCM 5 years ago may seem idealized today. An older model's use today can be justified as a concious effort to omit the impact of processes introduced in more recent models.

4.4 Boundary Conditions: A component of the diabatic hierarchy

Thus far we have focused on the question of how models represent internal diabatic processes; the representation of these processes are to some extent independent of the configuration of the models' boundary condition. Topography impacts the localization of storm tracks in the troposphere and plays a dominant role in the variability of the stratosphere (see Section 6.2). Topography may be included at all steps in the diabatic hierarchy. Surface energetics – either via prescribed SSTs or a slab ocean in more advanced models, or by simpler prescription of a direct heating in dry GCMs [e.g., *Chang*, 2006] – are a key consideration. When SSTs are prescribed, the surface effectively has an infinite reservoir of energy and the surface fluxes are unconstrained. This has the virtue of isolating mechanisms that operate via the heating of the atmosphere by convection and radiation. In Atmosphere Model Intercomparison Projects (AMIP) type experiments, SSTs are generally taken from observations or more comprehensive coupled climate model integrations. In Cloud Resolving Model (CRM) studies, SST is often uniform and constant, with the value serving as a key parameter in the experiment.

An alternative approach is to use a slab ocean, in which heat is exchanged with the atmosphere via radiative and turbulent surface fluxes and the SST is determined without explicitly modeling ocean advection [*Lee et al.*, 2008]. Ocean energy transport is a critical process to represent or prescribe, as many aspects of the climatological SST depend on it. Horizontal ocean heat transport can be included with prescribed flux of heat, often referred to as 'Q-fluxes' [e.g. *Russell et al.*, 1985].

Another key consideration is surface water availability and a widely used idealization is the aquaplanet, a water-covered globe without land. Aquaplanets have been broadly used since the 1980s, see references within *Neale and Hoskins* [2001], and more recently renewed efforts to catalog, and understand, differences between models in aquaplanet set-ups [*Blackburn et al.*, 2013; *Stevens and Bony*, 2013]. In aquaplanet set-ups the surface is usually assumed to be saturated with respect to liquid water vapor, with an infinite evaporation reservoir [*Neale and Hoskins*, 2001]. Aquaplanets can be run with both prescribed SSTs and with a slab ocean, and with zonally-uniform or zonally-varying SST patterns or ocean heat transports [*Shaw et al.*, 2015].

Some studies have extended the aquaplanet setup by including an idealized representation of land, with land being modeled as a very shallow ocean with reduced heat capacity, increased surface albedo and decreased surface evaporation [*Voigt et al.*, 2016; *Thomson and Vallis*, 2018]. Another commonly applied approach to represent the limited surface evaporation over land is the use of a 'bucket' surface hydrology [e.g., *Manabe*, 1969; *Byrne and O'Gorman*, 2013]. To isolate the impact of land-atmosphere coupling, one can also start from comprehensive land models and suppress the coupling by prescribing soil moisture [e.g., *Berg et al.*, 2016]. A few examples of idealized atmospheres with dynamic ocean models also exist [*Marshall et al.*, 2007; *Farneti and Vallis*, 2009], with the Farneti-Vallis model also including an idealized representation of land using a bucket model. By way of complement, there are also dynamic ocean models that couple to simple atmospheric model or prescribed atmospheric state, for example *Seager et al.* [1995]; *Deremble et al.* [2013].

The representation of atmospheric processes and the treatment of boundary conditions offer two components of the diabatic hierarchy. It is straightforward to use a single GCM with a hierarchy of boundary conditions, but generally less easy to build a hierarchy of atmospheric processes within a given model system. One system that is flexible to both approaches is the Weather Research and Forecasting (WRF) model, originally intended as a regional model. *Cesana et al.* [2017] showed that a similar range of cloud feedback uncertainty to that in CMIP5 can be generated in WRF by using different parameterizations. Another system that offers multiple parameterization options is the Isca framework [*Vallis et al.*, 2018a].

5 Mid-Latitude Circulation

The large-scale extratropical circulation provides one of the best success stories for hierarchical climate modeling: the dynamics of this circulation is now reasonably well understood and part of modern textbooks [e.g. *Vallis*, 2017]. Idealized simulations have played an instrumental role in this progress, providing key insights on the non-linear behavior of extratropical disturbances. Since the early days of climate modeling, theorists recognized the great power of numerical computing as a means to overcome the stringent limitations of analytical work. Idealized simulations aimed at understanding the atmosphere were performed in parallel with comprehensive simulations. Some of the insight gained with these early simulations constitute the basis of prevalent paradigms on the extratropical circulation.

We begin by highlighting two conceptual models that have allowed us to isolate the key elements of the midlatitude circulation and their interactions. The first is a class of *barotropic vorticity equation models*, where collapsing the vertical dimension allow us to focus on feedbacks between the zonal mean flow, Rossby waves, and the spherical geometry of the planet. The second is the *two-layer quasi-geostrophic channel model*, which provides perhaps the most simple context for understanding baroclinic instability. We then discuss two idealized modeling approaches that have been useful for studying the nonlinear baroclinic-barotropic dynamics in a simplified context: eddy life cycle experiments, and idealized forced-dissipative simulations.

5.1 Barotropic Dynamics on the Sphere

In addition to providing a the first numerical weather simulations [*Charney et al.*, 1950], the barotropic model served as a test bed to understand the influence of topography and localized heating on the general circulation [*Grose and Hoskins*, 1979; *Hoskins and Karoly*, 1981]. These experiments revealed the important role played by the mean flow structure for Rossby wave refraction in the upper troposphere. The widely used concepts of waveguides, propagation windows are based on these ideas, play a fundamental role for our understanding of the extratropical response to El Niño.

Despite our ability to now easily simulate the full three dimensional circulation, the so-called 'stirred' barotropic models [e.g., *Vallis et al.*, 2004] have seen a resurgence in recent years for understanding the dynamics of eddy momentum fluxes and eddy-driven jets without the complexity of baroclinic dynamics. The impact of baroclinic instability is, rather, approximated by a prescribed forcing (the stirring) of the vorticity equation at the synoptic scales. As a result, there are explicitly no feedbacks of the barotropic circulation on eddy generation.

The model has been used as a conceptual model of annular mode variability to explain the dependence of zonal index persistence on latitude [*Barnes et al.*, 2010] and to study the interaction between the tropical and subtropical jets [*O'Rourke and Vallis*, 2013], among other problems. As a further simplification, when the model is linearized it is possible to obtain a set of closed solutions (for simple forms of stirring) using stochastic theory [*DelSole*, 2001]. *Lorenz* [2014] has devised a very sophisticated method to calculate the eddy momentum flux given the full space-time characteristics of the stirring, which can play an important role due to the impact of wave phase speeds on refraction indices and wave propagation [*Barnes and Hartmann*, 2011]. The barotropic model can be a useful tool for exploring eddy momentum flux closures, which remain a challenging open question in general circulation theory and will be discussed in greater detail in Section 5.4.

5.2 The two-layer quasi-geostrophic model: conceptual model for baroclinic instability

To capture the essence of the baroclinic process, the *two-layer quasi-geostrophic model on the β-plane* stands out as a benchmark, indeed classical, model [*Phillips*, 1956]. It vies with the Eady model [*Eady*, 1949] as the simplest one that can produce baroclinic instability, albeit in a highly simplified form. There is only one baroclinic mode and the stratification and radius of deformation are prescribed, and the β-plane approximation and constant deformation radius make meridional propagation simpler than in the spherical case — the symmetry of the model makes northward and southward propagation equally likely. The model has been formulated in various configurations, forced or unforced, in doubly-periodic or channel domains. A popular setting is a forced-dissipative configuration, in which the model is forced by thermal relaxation to a baroclinic jet and the lower layer wind is damped using Rayleigh friction [*Zurita-Gotor*, 2007].

The two-layer model not only reproduces qualitatively the main features of the observed extratropical circulation but it also captures more subtle aspects of extratropical dynamics like the clustering of eddies in wavepackets [*Lee and Held*, 1993], the driving of low-frequency baroclinicity variability [*Zurita-Gotor et al.*, 2014] or the character of lower-troposphere eddy momentum fluxes [*Lutsko et al.*, 2017]. In a sense, this model is complementary to the barotropic model in that it is devoid of the barotropic feedbacks associated with sphericity that play an important role in the dynamics of that model. In the standard setting of the two-layer model, symmetry constrains the mean jet to be located at mid-channel but even when the jet does move (for example, when breaking the symmetry with a torque) these shifts have limited dynamical impact due to the plane geometry and use of constant f_0 and β.

5.3 The eddy life cycle paradigm

As a key simplification to the full non-linear problem, we cite the series of experiments systematized by Hoskins and collaborators in the 1970's, building on pioneering numerical work by *Edelmann* [1963] and others. The analysis of an eddy lifecycle (an initial-value problem for baroclinic instability) by *Simmons and Hoskins* [1978] introduced the notions of baroclinic growth and barotropic decay as an idealized conceptual model for the nonlinear evolution of extratropical disturbances. This simple paradigm has survived to today and plays a fundamental role for our understanding of wave-mean flow interaction and the maintenance of the mean circulation. Additional analysis [*Simmons and Hoskins*, 1980] uncovered the sensitivity of the decay stage in the lifecycle to the mean state, identifying two distinct patterns of evolution.

As theoretical advancements clarified the relation between eddy propagation and wave-mean flow interaction [*Andrews and McIntyre*, 1978; *Edmon et al.*, 1980] and the focus on PV dynamics highlighted the important role of wave breaking [*McIntyre and Palmer*, 1983], *Thorncroft et al.* [1993] proposed a conceptual model for understanding the two idealized lifecycles based on the direction of propagation and the typology of wave breaking. This is a very useful paradigm for understanding the dynamics of jet shifts and phenomena like the North Atlantic Oscillation [*Riviere and Orlanski*, 2007]. Idealized simulations were also useful for demonstrating the relevance of critical layer theory for eddy dissipation and wave-mean flow interaction in eddy lifecycles [*Feldstein and Held*, 1989]. The critical layer is a powerful concept for constraining upper-troposphere propagation [*Randel and Held*, 1991] and plays an important role for extratropical variability and climate sensitivity [*Lee et al.*, 2007; *Chen and Held*, 2007; *Ceppi et al.*, 2013].

5.4 Eddy closures and the sensitivity of the extratropical circulation

As the extratropical circulation is dominated by eddies, the key issue in extratropical modeling is determining the sensitivity of the eddy fluxes to changes in the mean state, the so-called closure problem. This has more than theoretical interest: although baroclinic eddies are well resolved by current models, the sensitivity of the eddy fluxes mediates the wide range of circulation responses to increased greenhouse emission across CMIP5 simulations [*Vallis et al.*, 2015]. A useful 'laboratory' for studying the closure problem is provided by *forced-dissipative models*, in which the mean-state is determined by the competition between the eddy fluxes and very idealized forms of forcing. These models can be formulated at different levels of complexity along the dynamical hierarchy depending on the scientific problem of interest.

The eddy heat fluxes determine the strength of the energy cycle and, in conjunction with the heating, the mean temperature gradient and the extratropical stratification. It is useful to decompose this problem in two parts: determining the meridional heat fluxes/temperature gradients given the stratification (a QG approach), and determining the stratification (and tropopause height) based on the column heat balance when the horizontal heat convergence is known [*Held*, 1982]. As the simplest conceptual model of the baroclinic extratropical circulation, the two-layer QG model (section 5.2) has played a prominent role in the development of eddy-mean flow closures with prescribed stratification [*Stone*, 1978; *Held and Larichev*, 1996]. These different theories have been tested in a hierarchy of forced-dissipative models at different levels of complexity, both with fixed and varying stratification [*Schneider*, 2004; *Zurita-Gotor*, 2007; *Zurita-Gotor and Vallis*, 2009; *Jansen and Ferrari*, 2013]. Although the impact of moisture on the extratropical circulation is still poorly understood, the association between available potential energy and stormtrack strength suggested by the simple models appears to explain the sensitivity of the extratropical stormtracks [*O'Gorman*, 2010]. Another question that may be relevant in a climate change context is the sensitivity of the extratropical circulation to the vertical structure of baroclinicity [*Butler et al.*, 2010; *Yuval and Kaspi*, 2016].

We expect the net eddy momentum convergence to scale with the strength of the energy cycle. A more subtle question is what controls the asymmetry in the eddy momentum flux, which has important implications for the jet latitude. In a QG beta-plane model the eddies propagate equally northward and southward of the stirring so the latitude of the jet is simply the latitude of maximum stirring. But on the sphere, equatorward propagation and poleward momentum fluxes dominate [*Thorncroft et al.*, 1993; *Balasubramanian and Garner*, 1997] so that we might expect extratropical jets to shift poleward as they strengthen if the stirring does not move. Additionally, idealized studies show the asymmetry between equatorward and poleward propagation to be sensitive to the latitude and scale of the eddies, barotropic shear and low-level baroclinicity [*Simmons and Hoskins*, 1980; *Hartmann and Zuercher*, 1998; *Rivière*, 2009], among other factors. As a result, understanding the dynamics of the poleward stormtrack shift with warming and its large inter-model variability remains a major challenge in climate theory.

5.5 Eddy feedbacks and the variability of the jet stream

To illustrate the use of hierarchical modeling in the extratropics, we discuss its application to the analysis of eddy feedbacks in unforced jet variability. We have chosen this example because it lends itself well to the hierarchical approach and because it is a topic of current research.

The leading (and more persistent) mode of extratropical zonal wind variability consists of a meridional shift of the eddy-driven jet concomitant with annular mode variability [*Thompson and Wallace*, 2000]. *Lorenz and Hartmann* [2001] found a positive correlation between the jet anomalies and their eddy momentum driving in the Southern Hemisphere when the jet leads by a few days (Fig. 3a), which implies that the anomalous eddy momentum fluxes tend to extend the duration of the jet anomalies. They interpreted this positive correlation as depicting the sensitivity of the anomalous eddy momentum flux on the state of the jet, or a positive eddy feedback (but see *Byrne et al.* [2016] for an alternative interpretation).

Climate models are known to be too persistent [*Gerber et al.*, 2008, see Fig. 3b], particularly idealized models [*Gerber and Vallis*, 2007]. This is mostly associated with too slow decay of the autocorrelation function at lags beyond 5 days (Fig. 3c), suggesting an excessive eddy feedback. Two different types of mechanisms have been proposed in the literature for this feedback: barotropic and baroclinic. Barotropic mechanisms rely on changes in upper-troposphere propagation due to changes in refraction in the presence of the anomalous jet, which may involve a number of different mechanisms [*Lorenz*, 2014; *Burrows et al.*, 2017]. In contrast, baroclinic mechanisms attribute the eddy momentum flux changes to changes in the stirring driven by the changes in the barotropic flow [*Robinson*, 2000].

Idealized models provide a useful framework for studying these two aspects of the problem in isolation. Using the stirred barotropic model, *Barnes et al.* [2010] investigated the sensitivity of the eddy momentum fluxes to the anomalous jet with fixed stirring. They showed that on the sphere, the eddy momentum flux becomes more asymmetric (equatorward propagation is enhanced) when the jet moves poleward, leading to a positive feedback. This may be understood in terms of changes in the turning latitude/reflecting level [*Lorenz*, 2014]. In the opposite direction, *Zurita-Gotor et al.* [2014] analyzed the dynamics of jet variability in idealized two-layer QG simulations and showed that the enhanced persistence in that model was consistent with the baroclinic feedback mechanism of *Robinson* [2000]. They found evidence of baroclinicity driving by the barotropic flow and very large coherence between the eddy heat and momentum fluxes at low frequency, with the momentum fluxes leading the variability (Fig. 3e). The co-variability between the barotropic and baroclinic components of the wind is also a robust result in observations [*Blanco-Fuentes and Zurita-Gotor*, 2011] and comprehensive climate models. Fig. 3d shows large correlation between the long-lag decay rates of (barotropic) jet anomalies and baroclinicity in a selection of CMIP5 models, so that models with more persistent jet variability also tend to have more persistent baroclinicity.

Stirred barotropic models can capture some aspects of the observed jet variability, like the persistence sensitivity to latitude [*Barnes et al.*, 2010]. On the other hand, the baroclinic mechanism may help explain the excessive persistence bias in comprehensive climate models (which cannot be corrected eliminating the jet latitude bias, *Simpson et al.* [2013]) or in idealized baroclinic models. Finally, diabatic effects may also play a role for annular mode persistence [*Xia and Chang*, 2014]. The jet persistence problem underscores the importance of making connections across the full model hierarchy, as the mechanisms at work may not be the same in all steps of the hierarchy, in comprehensive climate models and in the real atmosphere.

6 Middle Atmosphere Circulation

The middle atmosphere was initially a bit of an orphan in atmospheric research. Early work by *Charney and Drazin* [1961], among others, revealed that synoptic scale waves are generally trapped in the troposphere by lower stratospheric winds, such that early weather prediction studies could do reasonably well without a well-represented stratosphere. Higher above, the ionosphere plays a key role in telecommunications, necessitating research on the upper atmosphere. In recent decades, however, the representation of the troposphere has become sufficiently advanced that the importance of middle atmosphere processes can be appreciated in both weather prediction – particularly on subseasonal to seasonal time scales – and climate research [e.g., *Gerber et al.*, 2012].

In particular, the formation of the Antarctic ozone hole played a critical role in bringing stratospheric research to the fore. Despite the shorter history of middle atmospheric research, a number of key developments in atmospheric research were inspired by this region. Notably, wave-mean flow theory was developed in large part to explain and understand the circulation of the stratosphere. Lying above the tops of convection and clouds, the stratosphere is perhaps closest to the dry equation dynamics of theoreticians; one cannot explain the zeroth order circulation without understanding the essential role of waves in the transport of momentum, mass, and tracers. In addition, the essential roles of transport and chemistry in the formation of the ozone hole spurred research in these areas.

Here we highlight three conceptual models that have shaped our understanding of the stratosphere, and the more sophisticated steps in the hierarchy they have inspired. We first show how a single column model illustrates both the existence of the stratosphere and the response of the tropopause to global warming. The second example, the *Holton-Mass* model of wave mean flow interaction, provides a conceptual framework for interactions between waves and mean flow, and perhaps the simplest model for the extreme variability of the polar stratosphere. Finally, ozone depletion necessitated an understanding of transport within the stratosphere, and the so-called *leaky pipe model* provides a simple context to understand these processes. The Quasi-Biennial Oscillation provides another example of the advances that a simplified system can bring about, with the work of Lindzen, Holton and Plumb [*Lindzen and Holton*, 1968; *Holton and Lindzen*, 1972; *Plumb*, 1977] exposing the essential mechanism of the phenomenon years before GCMs could simulate it. Their work perhaps does not fit into a hierarchical framework so much as it provides a true 'theory' of the phenomenon, and thus the underpinning for parameterizations used in full GCMs, and we do not describe it here.

6.1 A 'single column' atmosphere and the response to global warming

A single column model of radiative-convective equilibrium (see also Section 8.1) allows one to explain both the existence of the stratosphere, as distinct from the troposphere below, and the basic vertical response of the atmosphere to global warming [e.g., *Manabe and Strickler*, 1964]. In pure radiative equilibrium — the balance between radiative cooling and radiative heating when there is explicitly no heat transport by the atmosphere — the warmest temperatures are found right at the surface, where the bulk of incoming solar radiation is absorbed. The radiative equilibrium solution, however, is unstable to convection, which will transport heat upwards to stabilize the profile and lead to a new equilibrium.

Convection produces a neutrally stratified layer up to a point where the 'radiative-convective equilibrium' matches the 'radiative equilibrium' profile. This matching point is guaranteed to occur because the radiative equilibrium profile becomes less steep with height, and in the case of an optically thin stratosphere that is transparent to incoming solar radiation the radiative equilibrium asymptotes to an isothermal profile with height. On Earth, the radiative equilibrium actually begins to increase with height at upper levels due to absorption of shortwave radiation by ozone. A single column model will thus naturally produce a sharp tropopause delineating two regions, a troposphere below where temperature is set by radiative-convective equilibrium and a stratosphere above where temperature is set by radiative equilibrium.

The single column model helps us understand the lifting of the tropopause [*Manabe and Wetherald*, 1967], who built on *Manabe and Strickler* [1964] by accounting for changes in water vapor (assuming constant relative humidity) accompanying increased CO_2 forcing. It is one of the most basic and robustly simulated responses of the atmosphere to greenhouse gas forcing [*Santer et al.*, 2003; *Vallis et al.*, 2015]. A related response of the stratosphere to greenhouse gas forcing is an increase in circulation, as quantified by a strengthening of the residual mean mass transport across pressure surfaces [*Butchart and Scaife*, 2001]. *Oberländer-Hayn et al.* [2016] show that net mass transport across the tropopause remains roughly constant, such that the tropopause provides a convenient measure of the lifting of the entire mass circulation [c.f., *Singh and OâĂŹGorman*, 2012]. Consequently, one can estimate changes in mass transport in the upper troposphere and lower stratosphere based on a prediction of the tropopause height. As the climatological mass transport decays sharply with height, a lifting of the circulation will lead to an increase in transport across pressure levels in the vicinity of the tropopause, as found by nearly all models [e.g. *Butchart*, 2014].

6.2 Conceptual models of stratospheric variability

The winter stratosphere in both hemispheres is dominated by a strong polar vortex, which effectively filters out synoptic scale variability from the stratosphere [*Charney and Drazin*, 1961]. In the early 1950s, however it was observed that the Northern Hemisphere vortex aperiodically undergoes a rapid breakdown, known as a Sudden Stratospheric Warming, or SSW [*Scherhag*, 1952]. *Matsuno* [1971] proposed a dynamical model of the warming based on the interaction of planetary waves propagating up from the troposphere that captured the basic mechanism, and later *Holton and Mass* [1976] developed a simple, essentially stratosphere-only, model that captured the seemingly oscillatory behavior of the vortex. The Holton & Mass study used a truncated baroclinic quasi-geostrophic model in which the wavenumber is constrained. The mean state was forced by Newtonian relaxation toward a specified state of radiative equilibrium, while the wave was forced by specifying its amplitude on the bottom boundary. This captured the essence of the stratospheric wave-mean flow interaction, with a transition between subcritical and supercritical behavior. In the supercritical regime, the wave grows, and the westerlies are weakened and even reversed: a prototypical SSW. This model is quasi-linear, in that wave-wave interactions are not represented. The model has continued to inspire research on the role of gravity waves in SSWs [*Albers and Birner*, 2014], and the role of the stratosphere on regulating wave activity [*Sjoberg and Birner*, 2014].

Multiple flow equilibria have also been demonstrated in more complex 3-dimensional stratosphere-only models, in which arbitrary height and latitude structure are permitted for the zonal flow and the waves [e.g., *Scott and Haynes*, 2000; *Scott and Polvani*, 2006], as opposed to using a single mode to represent the latitude structure. These stratosphere-only models provide a starting point for our understanding of wave-mean flow interaction and internal vacillations in the stratosphere; even with fixed lower boundary conditions, the stratosphere can exhibit substantial variability. *Scott and Haynes* [1998] also used a stratosphere-only model to suggest the possibility that the longer 'memory' of stratospheric winds in the tropics has an impact on extratropical stratospheric circulation.

An alternative approach that has been used is a 2-dimensional (horizontal) model in which a patch of vorticity is used to represent the vortex [e.g. *Polvani and Plumb*, 1992; *Esler and Scott*, 2005; *Esler et al.*, 2006]. These explored the relationship between the strength of topographic forcing and the propagation of Rossby waves on the edge of the vorticity patch, and narrowed the range of conditions which could result in a disturbed vortex. These studies provided some support for theoretical ideas on the role of resonance for stratospheric sudden warmings, particularly those in which the polar vortex splits.

Perturbations to the stratospheric vortex, either naturally through a SSW [*Baldwin and Dunkerton*, 2001], or by anthropogenic induced ozone loss [*Thompson and Solomon*, 2002] impact the tropospheric circulation. A stronger polar vortex in the stratosphere shifts the tropospheric jet streams poleward, and vice versa. A model that has been extensively used for studying the mechanisms of stratosphere-troposphere interactions in many different contexts has been that of *Polvani and Kushner* [2002]. This model simply extends the Held-Suarez model configuration to include a more realistic stratosphere, with a single parameter (the stratospheric lapse rate) controlling the strength of the winter polar vortex.

This relatively simple model demonstrated robust tropospheric responses to changes in the strength of the polar vortex. Subsequent modifications of this model have included extensions to study the seasonal cycle and better capture the structure of the lower stratosphere [*Jucker et al.*, 2013], and variations to study the impacts of tropospheric planetary-scale waves on stratospheric circulation and stratosphere-troposphere interactions in perpetual winter and with seasonality [*Gerber and Polvani*, 2009; *Sheshadri et al.*, 2015]. Similar studies were also carried out by *Taguchi et al.* [2001]; *Taguchi and Yoden* [2002] independently of the Polvani–Kushner model. These studies have employed more of a bottom-up than a top-down approach, in that they build up a simplified model containing just the essential ingredients for the study of the system of interest. An alternative approach would be to selectively simplify an existing GCM. Examples of such an approach include specified-dynamics and specified-chemistry version of full GCMs in attempts to separate the roles of interactive chemistry and dynamics in temperature and circulation changes [*Nowack et al.*, 2015; *Marsh et al.*, 2016; *Chiodo and Polvani*, 2016].

6.3 Conceptual models of stratospheric transport

Transport and chemistry play key roles in the distribution of trace gases throughout the stratosphere, including water vapor, ozone, and the substances that deplete ozone. Trace gases are both advected along by the mean Lagrangian circulation of mass, but can also be mixed along isentropic surfaces in the process of wave breaking. Such mixing leads to no net transport of mass, but will transport a trace gas if there is a horizontal gradient in its concentration. Early studies [*Holton*, 1986; *Mahlman et al.*, 1986] observed that tracers with very different chemical sources and sinks tend to exhibit common isopleths (surfaces of constant concentration).

These common features were explained by the 'age' of the stratospheric air, a construct designed to help untangle the roles of mass transport compared to mixing [*Hall and Plumb*; *Hall and Waugh*, 2000]. The mean age of a parcel is related to the mean time it takes for a parcel to travel from the surface of the atmosphere to any given location. In the troposphere, the age is on the order of hours (in convection) to days (baroclinic waves). In the stratosphere, however, the appropriate timescales are months to years. In practice, the age corresponds to an idealized tracer that increases with time (i.e. ages) in the free atmosphere, with the age set to zero at the surface (or, potentially, as it crosses the tropopause).

Since the age provides a measure of the time a parcel has been in the stratosphere, it quantifies how long chemical processes have been able to act. Its structure then mirrors that of any tracer whose source or sink is primarily in the troposphere, stratosphere, or upper atmosphere. For example, a tracer with a source in the stratosphere (e.g., ozone) will increase with the age, while a tracer with a sink (e.g., carbon monoxide) will decrease: they share isopleths despite exhibiting opposite gradients.

Early efforts to understand stratospheric transport examined limiting cases in the balance between transport of tracers across isentropic surfaces by the mean overturning mass circulation vs. the mixing of tracers along isentropic surfaces. *Plumb and Ko* [1992] consider a circulation where mixing along isentropic surfaces is extremely efficient. In contrast, *Plumb* [1996] developed the idea of a 'tropical pipe', where upwelling air in the tropics is entirely isolated from the downwelling air in the higher latitudes. Here, the age is set by the mean mass circulation alone.

These two limiting cases were combined in a benchmark model in our understanding of transport processes, the 'leaky pipe' model of *Neu and Plumb* [1999]. Following *Plumb* [1996], the leaky pipe divides the stratosphere into two regions, an upwelling 'pipe' in the tropics, and downwelling piped in the extratropics of both hemispheres. (In reality the two downwelling regions are entirely distinct, but the model seeks to capture their climatological distribution.) Mass is advected up the tropical pipe by the Lagrangian mean circulation (which can be quantified with the transformed Eulerian Mean or diabatic circulation), detraining continually out to the extratropics.

The boundary between the pipes, the edge of the surface zone, is a barrier to transport, but the 'leaky' pipe allows for some mixing of mass between the two. The most important parameters end up being the net detrainment and net mixing as a function of height, and can be solved analytically with appropriate simplifying assumptions. A key result of the model is that an increase in the net Lagrangian mass transport will tend to make the air younger, while a net increase in mixing tends to make the air older, as mixing leads to recirculation of air through the pipe.

While designed primarily as a conceptual model, the leaky pipe has been applied in a more realistic context to understand the make up of the stratosphere, and its response to anthropogenic forcing. *Garny et al.* [2014] use it to interpret changes in the stratospheric circulation in comprehensive models, separating the roles of mixing from the mean Brewer-Dobson Circulation. *Ray et al.* [2010] build on the leaky pipe to explain the distribution of trace gases, and *Linz et al.* [2016, 2017] use it to quantify the strength of the Brewer-Dobson Circulation from satellite measurements.

Early two-dimensional (latitude-height) transport models parameterized the impact of mixing before the full three-dimensional circulation could be resolved. Transport is affected by numerical diffusion (in addition to truncation errors associated with resolution), and continues to provide a limiting factor to our ability to properly represent stratospheric chemistry [*Karpechko et al.*, 2013]. Most climate and weather prediction models are run with 'specified chemistry', in which key radiative variables such as ozone are specified. Complementary to this, in Chemical Transport Models (CTMs) the dynamical variables are specified (typically from a reanalysis) to let the model simulate the transport and chemistry along the prescribed dynamic pathways. Integrations coupling chemistry with the circulation, so-called Chemistry Climate Models, are being investigated in the CMIP6 through the AerChemMIP [*Collins et al.*, 2017].

7 Tropical Circulation

Circulation and diabatic processes are intimately coupled in the tropics. A key scientific challenge has been to deconvolve the tight coupling between circulation, moisture, clouds, and convection. We focus in this section on conceptual models of the tropical circulation in which these processes (or their impact in the mean circulation) are prescribed, and defer the study of convections and clouds to Sections 8 and 9, respectively. Similar to the conceptual models of the mid-latitude circulation, many simple models for the tropical circulation hinge on reducing the dimension of the atmospheric flow. This can be done by vertical truncation, leading, for example to the Matsuno–Gill model for quasi-steady circulations. It can also be done by horizontal truncation, derived from truncation of the large-scale dynamics as in the Weak Temperature Gradient (WTG) approximation and related methods.

7.1 Vertical truncation: Matsuno–Gill and quasi-equilibrium conceptual models

As in other aspects of the model hierarchy, a canonical simplification is to vertically truncate the fluid governing equations. To this end, the Matsuno–Gill model utilizes the shallow water equations on an equatorial-beta plane with a single vertical layer of fluid, forced by prescribed heating [e.g., *Vallis*, 2017, section 8.5]. This framework may be interpreted in a number of ways, but it is most often thought to describe the horizontal structure associated with the first baroclinic mode with some equivalent depth. The first baroclinic mode is the most dominant, associated with latent heating (and so vertical motion) in the mid-troposphere and opposite signed horizontal flow in the upper vs. lower troposphere.

The shallow water equatinos allo for a detailed theoretical exploration of equatorial waves [*Matsuno*, 1966]. The Matsuno–Gill model can also describe the Walker circulation of the tropical Pacific and aspects of regional trade winds with prescribed latent heating *Gill* [1980]. The Matsuno–Gill model was also used as the atmospheric component of the first successful numerical ENSO prediction [*Cane et al.*, 1986] and the MJO, as detailed in Section 8.3.

There is an intimate link between the atmospheric circulation and latent heating in the tropics. Ascending vertical motion produces latent heat release, which, in turn, generates positive buoyancy that aids ascent. This link challenges 'dry thinking' in the tropics where heating is prescribed [e.g., *Gill*, 1980]. One cannot specify the heating structure to solve for the flow, as its structure *results* from the flow it is meant to describe [*Emanuel et al.*, 1994]. The full, interacting, moist system is then very complex, and (given the small scale of the convection) cannot be fully described by anything less than a cloud resolving model. Simplifications can and must be made, and the notion of convective quasi-equilibrium [*Betts*, 1973; *Arakawa and Schubert*, 1974] has become very influential. As well as forming the basis of many convection parameterization schemes used in GCMs with full vertical resolution, the model leads to a class of vertically truncated moist models. In these models convection consumes the potential energy at approximately the same rate that it is generated by large-scale processes, and the temperature profile in convective regions is constrained to be close to moist adiabatic. The Quasi-equilibrium Tropical Circulation Model (QTCM) [*Neelin and Zeng*, 2000; *Zeng et al.*, 2000] is one example of this approach. The QTCM was derived by considering the moist energetics with first-baroclinic vertical structure (e.g., as diagnosed in GCMs) and has played an important role for understanding and interpreting climate change projections [e.g., for radiatively forced precipitation changes *Chou and Neelin*, 2004; *Neelin*, 2007]. The model provided energetic perspectives on precipitation changes that were initially described and subsequently evaluated in GCMs [*Chou and Neelin*, 2004; *Chou et al.*, 2009]. A thorough evaluation of simple models of the quasi-steady tropical circulation is provided in the review of *Sobel* [2007].

7.2 Horizontal truncation: weak temperature gradient approximation

In the tropical free troposphere, horizontal gradients in pressure, density and temperature are small due to the smallness of the Coriolis parameter. Thus both horizontal advection and (on time scales longer than a day or so) the temperature tendency due to convection are small, and the dominant balance in the dry thermodynamic equation is between diabatic processes and adiabatic advection of potential temperature or dry static energy by the large-scale vertical motion. By assuming that this balance holds, so that the temperature equation becomes diagnostic for the large-scale vertical velocity rather than prognostic for temperature, one obtains a truncation of the large-scale dynamics that is often now described as the weak temperature gradient "WTG" approximation [*Sobel*, 2002], although the essential idea long predates that term, with origins in *Charney* [1963]. Methods that solve the same problem in different ways include the weak pressure gradient (WPG) [*Romps*, 2012a,b] and the very similar damped gravity wave method [*Kuang*, 2008]. WTG assumes that gravity waves efficiently homogenize the density distribution near the equator, where density anomalies cannot be rotationally balanced due to the small Coriolis parameter [*Charney*, 1963; *Sobel et al.*, 2001]. This approach is conceptually similar to the dynamical truncations in the mid-latitudes, where departures from balanced flows are used as the basis for reduced models of the flow, in that both assume gravity waves are fast compared to a slower, resolved component of the flow.

The WTG approach allows a representation of the large-scale tropical atmospheric circulations in explicitly convection permitting simulations on small horizontal domains. The large-scale vertical motion is interactive with the convection in the domain, so that neither one need be specified a priori — the model itself chooses whether to rain and how hard. In these simulations, one can perturb the surface conditions (e.g., SST, or surface wind speed) while holding the domain-averaged free-tropospheric temperature unchanged (or approximately so) to examine the response of the large-scale vertical velocity and precipitation [*Raymond and Zeng*, 2005; *Wang and Sobel*, 2011]. SSTs warmer than those that would be in convective quasi-equilibrium with the free-tropospheric temperature will provoke strong convection, large-scale ascent and adiabatic cooling to balance the associated diabatic heating, and an implicit water vapor convergence into the column that results in higher precipitation rates.

In Figure 4 the precipitation rate for such simulations using weak temperature gradient (top) and damped gravity wave (bottom) techniques are shown across the process hierarchy of resolved vs. parameterized convection (left to right). These simulations reproduce aspects of the observed relationship between column water vapor, large-scale vertical motion and precipitation [*Bretherton et al.*, 2004] with reduced dynamical complexity. WTG simulations of this type, with either cloud-resolving or single-column models, have also been used to understand the relationship between tropical drought and ENSO [*Chiang and Sobel*, 2002], tropical cyclogenesis [*Raymond et al.*, 2014], the sensitivity of tropical cyclone potential intensity to sea surface temperature [*Ramsay and Sobel*, 2011; *Emanuel and Sobel*, 2013] etc.

8 Tropical Convection

Convection spans a broad range of scales, from millimeters within boundary layers to global, as with the Hadley circulation. Moist convection describes areas of warm moist air that rise, condense, form clouds, mix with surrounding air and potentially rain. More formally, atmospheric moist convection describes thermally direct turbulent motions below the mesoscale (< 100 km) that result from vertical density perturbations.

While convection is an interesting phenomenon in its own right, a focus in the community has been how to represent it in GCMs, where grid cells are much larger than the mesoscale and so convection must be parameterized. Convection schemes fundamentally estimate tendencies (changes in time) of moisture and temperature in a grid cell due to the unresolved processes. The convection scheme determines if deep or shallow convection will occur (trigger function), the nature of the convective motions or effects as a function of height (cloud model), and the amount of convection (closure). Detailed reviews of convection schemes, and their history, can be found in *Emanuel and Raymond* [1993], *Stensrud* [2007] or *Plant and Yano* [2015] amongst others.

Convection is a very challenging process to model accurately across all space and time scales. A common view is that the explicit representation of convection will eliminate the need for parameterization in the foreseeable future, but this is far from assured given the very high resolution needed to fully resolve boundary-layer convective interactions; the important and difficult role of cloud microphysics (Section 9); and the observation that solutions do not always improve with resolution. Thus we believe a hierarchical approach is valuable to develop a better understanding of convection and its interaction with larger scales.

Models of convection range from fine-scale solutions of the full dynamical equations with interactive physics (large-eddy simulation or LES), to dry conceptual model such as Rayleigh-Bénard convection, to idealized plume or bubble models. We do not attempt to review all such models here; rather, we discuss some important idealized settings in which to study convection in order to improve our heuristic understanding of the phenomenon and how to represent its role at larger scales.

8.1 Simple configuration: radiative convective equilibrium

RCE is a paradigm for a statistical equilibrium of the Earth's climate [*Emanuel et al.*, 2014] and is one of the simplest forms of a quasi-equilibrium process. RCE was originally proposed as a simple framework for understanding global mean climate and its sensitivity to radiative forcing, and then evolved into a test-bed for understanding convection. RCE is a situation where radiative cooling to space balances heating generated by convection and time-invariant forcing results in a statistically steady state. RCE is an idealization that ignores the equator-to-pole temperature gradient and large-scale circulation. RCE is often taken as a simple approximation for the tropics valid for large space and time scales, but not locally due to the impact of the large-scale circulation [*Wing and Emanuel*, 2014].

RCE is an important component of a hierarchical approach connecting physical laws to the complex behaviour of the Earth system [*Popke et al.*, 2013]. RCE first appeared in the 1960s with *Manabe and Strickler* [1964] and for half a century since the RCE idealisation, has helped inform our understanding of convection. More recently, RCE has been used to understand convective organization in both CRMs and GCMs, where correctly reproducing convective organization in a start-of-art Earth system model remains problematic. organization describes convection that has a distinct structure, as opposed to being random, covering any scale from small-scale cloud clustering and squall lines, medium-scale such as mesoscale convective systems and the MJO, as well as global-scale such as the ITCZ. Here we focus on how the hierarchical approach is used to understand two forms of convective organization: self-aggregation in Section 8.2 and the MJO in Section 8.3.

8.2 Convective organization and self-aggregation

In CRMs without external forcing, i.e. homogeneous boundary conditions, convection spontaneously transitions from an initially homogeneous regime to a single convecting cluster, a process known as *self-aggregation*. The term aggregation more generally describes convection that is organized into clusters and is typically externally forced by circulation or temperature gradients. Self-aggregation depends on how clouds, radiation, convection, and the boundary layer are modeled [*Wing et al.*, 2017]. The complex interplay between model components make self-aggregation an excellent candidate for a hierarchical approach, although the mechanisms of aggregation remain to be fully determined.

We can not effectively review the broad literature on self-aggregation or convective organization more broadly. For more comprehensive analysis we refer the reader to: *Wing et al.* [2017] for self-aggregation, *Mapes* [2016] for a broader perspective, and *Holloway* [2017] for a comparison between observed and modelled aggregation. Instead we focus our discussion around how the hierarchical approach might be used to understand self-aggregation.

A benchmark model for studying self-aggregation is that of *Bretherton et al.* [2005]: a non-rotating three dimensional CRM in RCE with a constant sea surface temperature and no external forcing (see the top panel of Fig. 5). The horizontal domain size, $500 \times 500\mathrm{km}$, was large compared to previous studies. Their sufficiently large domain size appears to be a prerequisite for self-aggregation, so the smaller domains of earlier studies appear to account for why self-aggregation had not been seen observed previously. *Bretherton et al.* [2005] showed that limited organization occurred in the first 10 days but after 50 days a single cloud structure dominated the domain. In studying self-aggregation, a variety of approached for representing convection are used, including resolved convection in CRMs [*Bretherton et al.*, 2005] or global CRMs (GCRM) [*Satoh et al.*, 2016], parameterized convection [*Popke et al.*, 2013] or superparameterization [*Arnold and Randall*, 2015].

Self-aggregation is not solely a spatial reorganization of convection, but has dramatic impacts on the domain-mean climate in CRMs [*Wing et al.*, 2017]. Self-aggregation results in a very dry mean troposphere around the cloud clusters, more OLR in the domain mean, a warmer free troposphere and surface, decreased high cloud, increased low cloud, increased spatial variance of moist static energy and increased precipitation efficiency (see *Wing et al.* [2017] for a comprehensive list of references). While there is no domain-averaged vertical motion in RCE by definition, self-aggregation leads to circulation on the largest scales that are possible within the domain. The moist and dry sub-domains in CRMs with RCE are suggested to be two equilibrium steady states of a sub-system under WTG representing just those sub-domains. Two different equilibria, one dry and one moist, have been found in single column models (SCMs) [*Sobel et al.*, 2007] and small-domain CRMs under WTG [*Sessions et al.*, 2010] sharing the same temperature profile but different initial moisture fields, with the dry solution occurring for sufficiently dry initial conditions. The RCE and WTG simulations thus form their own hierarchy, with WTG helping to explain the phenomenon of self-aggregation occurring in RCE.

Self-aggregation is not a purely a CRM phenomena, but even occurs in GCMs. Using a non-rotating global aquaplanet with a coupled ocean and parameterized convection, *Popke et al.* [2013] showed convective clusters span broad regions, see their Fig. 2. Using a similar set-up but with prescribed SSTs *Becker et al.* [2017] found that self-aggregation, and global climate, are sensitive to the convective parameterizations. Using a collection of rotating AMIP simulations, *Maher et al.* [2018] showed that without parameterized convection, precipitation is more clustered on daily time-scales and extreme precipitation had twice the rain rate. Using a non-rotating GCM in RCE *Reed and Medeiros* [2016] applied the reduced planet approach – decreasing the planetary radius rather than increasing resolution – to show the transition of large scale aggregation through to CRM-like self-aggregation.

With the addition of planetary rotation, self-aggregation morphs into tropical cyclones, and their change in intensity and frequency with climate change is an area of great societal importance. The tropical genesis regions are over warm tropical oceans and so aquaplanet simulations are appropriate, typically using RCE with careful consideration in setting up meridional temperature differences in order to generate and maintain the tropical cyclones. These idealized configurations have aided in our understanding of tropical changes with increasing SSTs [*Held and Zhao*, 2008; *Khairoutdinov and Emanuel*, 2013; *Merlis et al.*, 2016] and tropical cyclone characteristics [*Shi and Bretherton*, 2014; *Satoh et al.*, 2016]. In the bottom panel of Fig. 5, the GCRM simulations of *Satoh et al.* [2016] are shown with and without rotation for both prescribed and observed SSTs.

Simplified though they may be, the above models of radiative-convective equilibrium and convective aggregation are still very complex compared to the convection models typically used in physics and turbulence studies (as, for example, described in *Chillà and Schumacher* [2012]) and which may also lead to pattern formation and aggregation. The literature in two fields (moist atmospheric convection and Rayleigh–Bénard-type convection) is almost completely non-overlapping yet the subjects themselves have much in common. Their connection has begun to be explored by *Pauluis and Schumacher* [2011] and *Vallis et al.* [2018b] with simple models that seek to capture the essential dynamics of moist convection and which may form the base of the atmospheric moist convection hierarchy, but this end of the hierarchy is still largely unexplored.

8.3 Convective organization: the Madden–Julian oscillation

The Madden–Julian oscillation (MJO) is an envelope of organized tropical convection that drifts eastward from the Indian Ocean into the Pacific. It is distinct from most convectively coupled equatorial waves in having a relatively slow speed of propagation ($\approx$ 4–8 m/s) and long timescales (about 1–2 months), and a relatively large scale (planetary wavenumbers 1–3) compared to other synoptic disturbances in the tropics. Our theoretical understanding of the mechanisms that initiate, propagate and maintain the MJO are incomplete [*Ahn et al.*, 2017]. While it has also been historically difficult to simulate in global models, some recent models do much better than past ones, and in fact some dynamical forecasts are now superior to statistical ones. This new simulation capability allows theoretical ideas to be tested. The MJO's complex interactions between moisture, clouds, radiation, convection and circulation make it an excellent candidate phenomena for hierarchical approach, with the comprehensive models anchoring the hierarchy.

Realistic simulations of the MJO require convection to be sensitive to free-tropospheric moisture, i.e. a positive moisture-convection feedback—free-tropospheric humidity is higher in regions of deep convection. CMIP5-class models with the largest moisture sensitivity tend to have the most realistic MJO [*Kim et al.*, 2014a]. Poor simulations of the MJO — generally those with weak to non-existent MJOs [*Ahn et al.*, 2017] — can be improved by increasing the sensitivity of convection to moisture, such as increasing the entrainment and rain re-evaporation. Such tuning for the MJO generally causes biases in mean climate [e.g., *Kim et al.*, 2011], but there is some evidence to suggest a realistic MJO and mean state can occur simultaneously even with traditional convection schemes [*Crueger et al.*, 2013]. There is considerable additional evidence, apart from the MJO, that deep convection in general is quite sensitive to moisture [e.g., *Derbyshire et al.*, 2004] and that typical convective schemes have excessive undilute ascent, as opposed to entraining, [e.g., *Tokioka et al.*, 1988; *Kuang and Bretherton*, 2006].

More recent studies have viewed the MJO through the moist static energy budget where surface fluxes and radiation are the dominant source terms (since moist static energy is conserved under condensation, which is the dominant source term in the dry static energy budget in deep convective conditions). Feedbacks between surface turbulent fluxes and convection were emphasized in early theories [*Neelin et al.*, 1987; *Emanuel*, 1987] and appear to be important in some GCMs [e.g. *Maloney and Sobel*, 2004]. Other work, however, points to cloud-radiative feedbacks as more important—less longwave cooling by high-clouds in a moist atmosphere— in GCMs [*Andersen and Kuang*, 2012; *Chikira*, 2013], process-based diagnostics [*Kim et al.*, 2015] and so-called "mechanism denial" experiments [*Kim et al.*, 2012; *Crueger and Stevens*, 2015; *Ma and Kuang*, 2016]. This is consistent with earlier work with more idealized models: *Raymond* [2001] argued that radiative feedbacks were important to the MJO based on results from a 3D model of intermediate complexity, while *Bony and Emanuel* [2005] did so based on 2D CRM simulations without rotation and *Hu and Randall* [1994] found radiative feedbacks are critical in a one-dimensional model without large-scale circulation.

The importance of moisture-convection and cloud-radiative feedbacks suggests a view of the MJO as essentially a form of self-aggregation on the equatorial β-plane, in a domain much larger than CRMs simulations [e.g. *Arnold and Randall*, 2015]. In aquaplanet simulations with super-parameterized convection in RCE, *Arnold and Randall* [2015] found similar energy budgets and radiative feedbacks in non-rotating simulations, where self-aggregation dominates, and simulations with rotation, where MJO-like variability occurs.

The importance of moisture-convection and cloud-radiative feedbacks are the core assumptions in a recent set of highly idealized models of the MJO. These models represent the MJO as a moisture mode – a mode that would be absent in a dry atmosphere. In these idealized models, essential information is contained in the moisture field. Truncation to a single vertical mode, as in the Matsuno–Gill model, allows the dry dynamics to become shallow water-like. The convection schemes depend strongly, and in some cases exclusively on the moisture field, building in a strong moisture-convection feedback.

Moisture modes emerged in the idealized models of Fuchs and Raymond [*Fuchs and Raymond*, 2002, 2007; *Raymond and Fuchs*, 2007, 2009]. The moisture mode was isolated in the simple 1D linear model of *Sobel and Maloney* [2012, 2013] that has a single moisture prognostic variable, assumes WTG in the temperature equation, and generates winds by assuming a Matsuno–Gill response to quasi-steady heating (approximately valid as long as the disturbance does not propagate too quickly). In this model it can be shown explicitly that radiative feedbacks are critical for eastward propagation in a linearly unstable mode [*Sobel and Maloney*, 2013]. While the eastward propagation was initially slower than observations, modifications by *Adames and Kim* [2016] increased the propagation speed by accounting for meridional moisture advection. Because the WTG assumption eliminates the Kelvin waves, the waves that most early theories relied on to explain the eastward propagation, the propagation of a moisture mode results largely from horizontal moisture advection, which seems to be supported by a number of observational and modeling studies [e.g., *Maloney*, 2009; *Pritchard and Bretherton*, 2014; *Kim et al.*, 2014b; *Inoue and Back*, 2015a].

Moisture mode theory — including the link to self-aggregation in idealized simulations — provides a useful framework for diagnosing models and observations, although whether moisture mode models correctly capture the MJO remains a topic of debate. The moisture mode ideas are quite different from those in earlier MJO theories, most of which excluded both radiative feedbacks and prognostic moisture (e.g., see review by *Wang* [2005]), and also differ from other, more recent models [e.g., *Majda and Stechmann*, 2009; *Yang and Ingersoll*, 2013]. A connection to the moisture mode hypothesis is provided, however, through a hierarchical chain from self-aggregation in idealized simulations through to more realistic ones, where moisture-convection and radiative feedbacks are critical. As some comprehensive models have come to simulate the MJO with much greater fidelity than in the past, it is critical that theories of MJO behavior and comprehensive models make better connection to each other, with the latter used to test ideas through mechanism denial or other experiments designed for the purpose.

9 Clouds

Clouds span time scales from seconds to years and spatial scales from droplets to planetary waves, and the radiative impact of clouds is fundamental for modeling Earth's weather and climate. The broad range of scales and impacts suggests a hierarchical approach and various approaches, at different levels of complexity, are beginning to emerge. We focus our discussion around the importance of clouds in shaping the radiative properties of the atmosphere and its circulation, and refer the reader to *Schneider et al.* [2010] for a review of water phase changes and latent heating.

Clouds affect both the circulation and, more directly, the climate sensitivity, and one of the World Climate Researh Program's 'Grand Challenges' is centered on clouds, circulation and climate sensitivity [*Bony et al.*, 2015]. A primary cause of inter-model spread in climate model estimates of equilibrium climate sensitivity is the response of clouds to changes in external forcing. It has been known since the 1970s, when GCMs represented clouds only in a rudimentary manner, that clouds are a source of uncertainty [*Arakawa*, 1975; *Charney et al.*, 1979], and this cloud uncertainty has persisted through today even though cloud parameterizations have evolved [*Boucher et al.*, 2013] — a telling example that sophistication does not lead to convergence. The difficulties in representing clouds are myriad, but ultimately rooted in the small scales that are involved compared to the coarse resolution of global models. Reviews of these difficulties are provided by, e.g., *Arking* [1991]; *Stephens* [2005]; *Stevens* [2005] and *Ceppi et al.* [2017].

In the following two sections we look at both theses issues. We first focus on the coupling between clouds and large-scale circulation, and how it can be studied by manipulating cloud-radiative interactions in GCMs. Then, in Section 9.2, we address the impact of clouds on Earth's energy balance and focus on climate sensitivity.

9.1 Cloud-circulation coupling via radiation and the treatment of clouds

Many aspects of the general circulation can be simulated without clouds, but an understanding the cloud-circulation coupling, and in particular the coupling via radiation, is needed for a truly quantitative understanding of the circulation. The large-scale circulation and thermodynamic structure determine the bulk distribution of clouds, and the clouds then feedback onto the atmospheric state, but a proper understanding of this loop, and even whether the feedbacks involved are positive or negative, has proven elusive. The problem itself is been recognized for some time — the importance of the radiative effects of clouds on the circulation has been known since the 1980s [*Hunt et al.*, 1980; *Slingo and Slingo*, 1988; *Randall et al.*, 1989], although how these effects may change as the planet warms has come more to the fore of late [*Bony et al.*, 2015; *Voigt and Shaw*, 2015]. More detailed studies have followed, and the cloud-circulation coupling problem in the tropics was explored by [*Voigt et al.*, 2014; *Merlis*, 2015; *Feldl and Bordoni*, 2016; *Crueger and Stevens*, 2015] and extratropical problems and *Voigt and Shaw* [2015]; *Ceppi and Hartmann* [2016] and others have looked at extratropical issues.

The primary challenge in studying the radiative cloud-circulation coupling may be thought of as adapting the diabatic hierarchy to decouple cloud-radiative effects from the circulation. Various methods have been developed to accomplish this task, each of which try to isolate a pathway or a feedback: an atmospheric pathway results from the direct impact of cloud-radiative effects on the thermodynamics of the atmosphere in the absence of SST changes, and a surface pathway results from the impact of cloud-radiative effects on the surface energy balance and thus on SSTs.

Perhaps the simplest method is to force a dry GCM with atmospheric cloud-radiative effects simulated from GCMs [*Voigt and Shaw*, 2016]. A second method is to use intermediate GCMs without clouds, e.g., the gray-radiation aquaplanet of *Frierson* [2007]. Using Frierson's model, *Kang et al.* [2009] showed that the tropical circulation response to extratropical forcing is muted compared to comprehensive GCMs, emphasizing the fundamental role that clouds play in shaping the energetic need for a cross-equatorial Hadley circulation. A third method is to use cloud-locking [*Zhang et al.*, 2010; *Mauritsen et al.*, 2013; *Voigt et al.*, 2014] – using prescribed cloud fields to remove the radiative coupling. Because cloud radiative effects depend nonlinearly on cloud properties, constant time-mean clouds do not in general yield the correct time-mean radiative fluxes. It has thus proven useful to prescribe time-dependent clouds that vary with the time step of the GCM's radiation scheme. Cloud-locking simulations have for example suggested that clouds are fundamental in setting the ITCZ sensitivity to hemispheric perturbations [*Voigt et al.*, 2014]. Finally, the transparent-cloud method uses clear-sky instead of all-sky radiative heating rates [*Randall et al.*, 1989; *Merlis*, 2015; *Albern et al.*, 2017] and is easier to implement than the cloud-locking method. Transparent-cloud simulations have highlighted that cloud radiative effects strengthen the Hadley cell and eddy driven jet stream, reduce tropical-mean precipitation, and narrow the ITCZ [*Li et al.*, 2015; *Harrop and Hartmann*, 2015; *Popp and Silvers*, 2017; *Albern et al.*, 2017].

Still less is known about cloud-circulation coupling in the extratropics. For example, the fact that there are only limited observations over the Southern ocean may contribute to modeling biases in the Southern Hemisphere, such as the equatorward bias in the position of the eddy driven jet stream [*Kidston and Gerber*, 2010]. *Ceppi et al.* [2012] found that the eddy-driven jet bias results in part due to too little shortwave reflection from Southern ocean clouds. Cloud-radiative effects were also shown to be important for the modelled circulation response to global warming. Simulations with comprehensive GCMs (in idealized aquaplanet and more realistic settings) using the cloud-locking method have demonstrated that half or more of the extratropical circulation response to global warming can be attributed to radiative changes in clouds [*Voigt and Shaw*, 2015, 2016; *Ceppi and Hartmann*, 2016], and that both the atmosphere and surface pathways contribute to the cloud impact.

The use of the model hierarchies to instigate the role of regional cloud changes for the jet stream response to global warming is illustrated in Figure 6 [*Voigt and Shaw*, 2016]. Coupled GCMs show large differences in the jet stream response over the 21st century, in particular in the Southern Hemisphere (Fig. 6 a). These differences persist in idealized prescribed-SST aquaplanet simulations with the same models, indicating an important role of cloud-radiative changes and the atmospheric pathway (Fig. 6 b). In the MPI-ESM model cloud-radiative changes alone cause jet stream changes (Fig. 6 c) as large as the model-mean response, and arise mainly from changes in high-level tropical and mid-latitude clouds (colored lines). This response is also reproduced in the MPI-ESM dry Held-Suarez set-up (Fig. 6 d). In combining the cloud-locking method with different model setups the cloud-radiative impact on the projected extratropical circulation response to global warming is, we may hope, better understood.

9.2 Cloud-radiative feedbacks on climate sensitivity

Equilibrium climate sensitivity describes the global average change in surface temperature due to doubling CO_2 and is a useful and widely used measure of climate change. In spite of tremendous increases in climate model verisimilitude over the past 40 years, estimates of equilibrium climate sensitivity have remained the same, with state-of-the art models producing answers generally ranging from 1.5–4.5 K, with a few outliers having a larger sensitivity and with some clumping around 2.5–3 K. The primary source of the spread in this sensitivity, and hence in our uncertainty as to its true value, is cloud feedbacks which vary broadly across models [*Boucher et al.*, 2013; *Chung and Soden*, 2017]. And although the full panoply of feedbacks in comprehensive climate models is needed to determine climate sensitivity, atmosphere-only models, and indeed idealized atmosphere-only models in, for example aquaplanet configurations, may be better suited to isolate the cloud feedbacks [*Medeiros et al.*, 2008; *Ringer et al.*, 2014; *Medeiros et al.*, 2015].

Cloud feedbacks are often estimated as changes in the top of atmosphere cloud radiative effect – the difference in net radiative fluxes between clear-sky and all-sky conditions (i.e., the impact of clouds). A challenge in understanding cloud feedbacks is that clouds have both shortwave (visible/albedo) and longwave (infrared/greenhouse) effects. High-level cirrus clouds can have large radiative impact, but the shortwave and longwave response are of different sign and partially offset one-another, with a slight warming effect overall. On the other hand, low-level clouds emit longwave radiation at a temperature close the surface so their dominant effect is in the shortwave.

Consider first the high-level cloud feedbacks. There have been many hypotheses for cloud feedback mechanisms that have relied upon high-level clouds producing a negative (stabilizing) feedback [*Ramanathan and Collins*, 1991; *Lindzen et al.*, 2001] under global warming. These hypotheses, in their original form, have largely since been refuted (though recently revisited by *Mauritsen and Stevens* [2015]), who argue that substantial changes in large scale aggregation could have a cooling effect). A more robust hypothesis concerning the behavior of high clouds is the 'Fixed Anvil Temperature' (FAT) hypothesis [*Hartmann and Larson*, 2002; *Hartmann et al.*, 2001]. This hypothesis argues that anvil clouds in regions of tropical outflow will remain at approximately the same temperature as the surface warms, as they depend on the level of maximum divergence in cloud-free regions that radiatively cool [*Hartmann et al.*, 2001]. The FAT hypothesis seems to be generally well supported by numerical simulations, using both numerical weather prediction type models *Hartmann and Larson* [2002]; *Larson and Hartmann* [2003] and cloud resolving models [*Kuang and Hartmann*, 2007]. The FAT hypothesis successfully explains the robust positive feedback associated with longwave cloud radiative effects in comprehensive climate models [*Zelinka and Hartmann*, 2010]. and, more generally, may be regarded as a success story in the fraught area of cloud feedbacks.

We have much less success in understanding the shortwave feedbacks associated with low-level clouds, although there is mounting evidence that the feedback is positive [*Klein et al.*, 2017]. A hierarchical approach to understand the complex shortwave cloud feedback is appealing due to the success in developing the FAT hypothesis and because there are so many possible mechanisms and sources of uncertainty [*Bretherton*, 2015] that without simplification the task is hopeless. That the task is complex can be seen from the results of two studies. Using a hierarchy of models — including comprehensive, atmosphere only, aquaplanet, and single column configurations — *Brient and Bony* [2013] identified a positive feedback that depends on how moist static energy is transported between the free troposphere and the boundary layer. However, comparing SCMs configurations of several GCMs in idealized climate change experiments, *Zhang et al.* [2013] showed that different GCM physics still produced different cloud responses, suggesting that the treatments of shallow convection and boundary layer turbulence are key differences among models.

Two idealized model set-ups for studying clouds are SCMs and aquaplanets. SCMs are used to explore how parameterized physics can respond to climate perturbations [*Dal Gesso et al.*, 2015]. A limitation of this approach is that SCMs experiments are difficult to compare with GCM experiments where clouds and circulation are fully coupled. A couple of studies, focused more on convection than on clouds per se, have done this using WTG to represent the convection-circulation coupling and compare SCMs and GCMs solutions explicitly [*Raymond*, 2007; *Zhu and Sobel*, 2012]. Using aquaplanet and realistic topography configurations, *Medeiros et al.* [2008] found that the cloud response to prescribed SST warming are similar in each model set-up. The intrinsic value of the aquaplanet is that it removes complexities that may obscure fundamental underlying physics [*Stevens and Bony*, 2013]. In the case of cloud feedbacks, model comparisons continue to support the notion that parameterized physics associated with shallow convection are at the heart of uncertainty in estimates of equilibrium climate sensitivity [*Ringer et al.*, 2014; *Medeiros et al.*, 2015]. The symmetry associated with aquaplanets has also helped to emphasize the role that regional feedbacks play for climate sensitivity, in particular by pointing toward nonlinear feedback evolution [*Feldl and Roe*, 2013; *Rose et al.*, 2014; *Roe et al.*, 2015; *Andrews et al.*, 2015; *Zhou et al.*, 2016]. Aquaplanet GCMs in RCE are a further useful idealization. Investigating feedbacks and climate sensitivity in an

RCE configuration may further refine the scope of the problem by isolating tropical processes and focusing on the model physics [*Bony et al.*, 2016; *Popke et al.*, 2013]. Reducing complex GCMs to RCE configurations also makes direct contact with earlier theoretical work on climate feedbacks using RCE in a single-column setting [*Manabe and Strickler*, 1964].

10 Summary

We have highlighted a number of models that, by virtue of isolating key atmospheric processes, have found widespread use across the atmospheric sciences. Some have become 'benchmarks', by which we mean a standard to validate a new numerical code against, e.g., the *Held and Suarez* [1994] test for a new dynamical core, or a model that underpins our conceptual understanding, e.g. radiative-convective equilibrium as an abstraction of the tropical circulation. The models are diverse in nature, as illustrated in Figure 7 which shows the models available in CESM: the Earth system (fully coupled with an ocean), atmosphere only, aquaplanet, RCE, and dry physics integrations.

The principles of dynamics, process, and scale help us classify existing hierarchies (Section 2). From a practical standpoint, an atmospheric model can be viewed as a construction of these elements. One must choose the physics, the appropriate governing equations of fluid flow; the *forcing*, the processes regulating the thermodynamic and dissipative processes within the free atmosphere and boundaries; and the *scale* of the domain, the size, geometry, boundary conditions, and resolution. Model hierarchies are created along all three of these components, and many of the most useful hierarchies rearrange these basis functions to chart appropriate paths through this space.

The diabatic hierarchy of Section 4 highlights a family of models focused on two components, the representation of diabatic processes and the boundary conditions. Along the first component, we identify three steps in ascending order of complexity:

- Dry GCMs, where atmospheric thermodynamics are reduced to Newtonian relaxation to a specified equilibrium temperature,
- Idealized moist GCMS, where water vapor is a prognostic variable transporting latent heat, but where it's interaction with radiation is severely limited (enabling one to effectively remove microphysics), and
- Comprehensive AGCMs, which seek to represent the critical interactions of water (in all phases) with radiation.

The second component – the boundary conditions – can be applied (at least to some extent) at each level in this hierarchy, e.g. specified SSTs, slab-ocean, or coupled atmosphere-ocean, the treatment of 'land' (land-sea contrast, water availability, and topography), and, from above, the representation of the stratosphere, e.g., resolution of the tropopause region and treatment of subgrid scale gravity waves.

Sections 5-7 focus on our understanding of the atmospheric circulation, allowing us to consider even more fundamental models. The midlatitude circulation is governed by the evolution of synoptic scale eddies, and their fluxes of momentum and heat, both sensible and latent. In Section 5 we explored three conceptual models that provide a foundation for understanding the interactions between synoptic eddies and the large scale jet streams:

- Barotropic vorticity dynamics on the sphere, which isolate the feedback between the zonal flow and eddies on the eddy momentum fluxes that effect the extratropical jets and storm tracks,
- The two-layer quasi-geostrophic model in a channel, the simplest model to capture baroclinic and barotropic eddy interactions.
- The eddy lifecycle – an initial value approach to understanding eddy evolution, which can be run across a wide spectrum of models.

Coupled with the use of global models from the diabatic hierarchy, these benchmarks have enabled us to make progress with the closure problem (Section 5.4), and understand feedbacks with the persistence of the jet streams that have bedeviled comprehensive models (Section 5.5).

The ozone hole brought the stratosphere to forefront of research in the 1980s and 90s [and more recently, with the recognition that it plays a key role in observe circulation changes, e.g., *Polvani et al.*, 2011], requiring an understanding of dynamics, transport, and chemistry. The absence of latent heat transport and microphysics (polar stratospheric clouds excepted) also provided an ideal environment for the development of wave mean flow theory, which in turn fundamentally improved our understanding of the troposphere. Section 6 highlights three conceptual models that capture essential features of the stratosphere:

- Single column radiative-equilibrium, a starting point to understand formation of the tropopause and its response to global warming,
- The *Holton and Mass* [1976] model of wave mean flow interaction, capturing multiple equilibrium and the germ of a Sudden Stratospheric Warming, and
- The age of air [*Hall and Plumb*] and the leaky pipe model of *Neu and Plumb* [1999], the basis for our understanding of tracer transport through the stratosphere.

Chemistry adds another element to the diabatic hierarchy – allowing the radiative active species within the atmosphere to evolve dynamically – and the AerChemMIP [*Collins et al.*, 2017] is exploring this frontier as part of the CMIP6.

Looking forward, there has been an effort to cease viewing the middle atmosphere and troposphere as isolated systems, rather taking a holistic approach to understanding the circulation. The feedbacks controlling the persistence of natural variability in the troposphere discussed in 5.5 are influenced by the stratosphere. In dry dynamical cores (Section 4.1), changes only to equilibrium profile above 100 hPa impact both the position and persistence of the jet stream. The primitive equation dynamics still present a conceptual challenge, and there is a need for a conceptual model tying the tropospheric jets with the polar vortices.

In the tropics, the central role of moist processes and weak rotation (and so the inability to fall back on quasi-geostrophic scaling) presented a great challenge to modeling efforts. Global scale phenomenon, such as MJO, depend on small (km) scale convective and cloud processes, and hence still present a challenge to state-of-the-art models. None-the-less, Section 7 reviews the substantial progress has been made. We highlight three conceptual models:

- Matsuno–Gill models of the first baroclinic mode, a conceptual model used, for example, to understanding the atmospheric responses to local heating,
- Quasi-Equilibrium, an assumption on which some convection schemes are based where large-scale processes are balanced by convection, and
- The weak temperature gradient approximation, used to approximate tropical dynamics in order to study interactions with smaller-scale processes.

Convection in Sections 8 and clouds in Section 9 focus on the moist processes that must be parameterized in climate prediction models. However, identifying the benchmark models is less straightforward. For tropical convection we identify:

- Radiative Convective Equilibrium, a idealized model set-up that describes a balance between radiative cooling and convectively generated heating, has shaped our understanding of convective organization, in particular self-aggregation and the MJO, and
- Linear models of tropical baroclinic modes are used to understand the moisture model view of the MJO and the coupling of convection and large-scale circulation more generally.

While the essential physics of clouds are relatively well understood, faithfully representing clouds in simplified models (or, for that matter, complex models) has not been successful enough for consensus to emerge on a hierarchy of cloud models. Part of this lack of consensus is because of the strong interaction between clouds and radiation, making feedback processes important and hampering the utility of high-resolution simulations because of computational resource limits. Nevertheless, the following models have and, we believe, will continue to prove helpful:

- High-resolution simulations (cloud-resolving and large-eddy), which avoid many of the complications and compromises of parameterization, provide a process-level view of clouds and convection and a benchmark against which other models can be compared.
- Single-column models that remove the cloud-circulation feedback, but allow efficient exploration of column-based parameterized physics.
- Aquaplanet models (including radiative-convective equilibrium configurations) with both prescribed and interactive SSTs, which provide idealized investigation of cloud-radiative feedbacks and cloud-circulation coupling.
- Methods that short-circuit part of the cloud problem, e.g., the clouds-off and the locked-clouds approaches, and effectively demonstrate the effects of clouds.

11 Outlook

Growing computational resources consistently push the frontier of atmospheric research to more complex models, allowing us to run at higher resolution and account for new processes. With numerical weather prediction, the gains from enhanced observation and assimilation capacity and more sophisticated, higher resolution models have led to clear, measurable improvements in forecasts [e.g., *Bauer et al.*, 2015]. With respect to climate change, however, our predictions and uncertainty bounds have not changed much since the pioneering report of *Charney et al.* [1979]. We now account for more degrees of freedom in current models — by orders of magnitude — and understand much more about the details of the atmosphere and its role in climate. But in practical terms, our inability to narrow the confidence intervals has not helped policy makers.

Climate prediction differs from weather prediction in that improvements to our observational network do not immediately allow us to observe all the time scales relevant for climate. In addition, we must face the critical role of atmospheric physics that we cannot directly simulate – clouds, convection, chemistry, and microphysics, among others. Better models (and the larger computers we need to run them) will clearly be essential to improving projections of future climate. As the distance between our theoretical understanding and our most sophisticated models becomes greater, however, we argue that the need for a hierarchy of models will only become more important.

The remarkable ability of *Charney et al.* [1979] almost 40 years ago to estimate the climate sensitivity of our atmosphere – in possession of computers weaker than those in current mobile phones – is a testament to the value of deep theoretical insight. If comprehensive models are the frontline of our field then model hierarchies are the supply chain: the connection between what we can comprehend as a human, and what we can do with our technology.

Looking forward, we suggest a few areas where the model hierarchy could be expanded to enable progress. For example, there may be a role for models that bridge the gap between dry dynamical theory and the moist atmosphere. As discussed in sections 5–7, most of the conceptual models of the atmospheric circulation do not explicitly include moist processes. Our understanding of the extratropical circulation is largely based on dry theories which do not explicitly account for moisture, but nonetheless do a good job of explaining key aspects of the circulation's response to warming in comprehensive climate simulations [e.g. *Lu et al.*, 2007; *O'Gorman*, 2010]. Even in the tropics, however, the Matsuno–Gill conceptual framework does not explicitly include moist processes. Beyond some theoretical ideas which suggest that one could capture the impact of moisture in a dry model by using an equivalent (and reduced) dry stability [e.g., *Emanuel et al.*, 1987; *O'Gorman*, 2011], there remains a significant gap between dry and moist models in the diabatic hierarchy.

A second area is to include processes that dominate uncertainty in climate prediction – for example, aerosols and microphysics, or chemistry and transport – down into the hierarchy. These key processes generally only appear in our most complex models, where one must study them in concert with all other parameterizations. In addition to the practical limitations of using a state-of-the-art model (which can often only be done within a modeling center), the target is continuously moving with the march of the CMIP and the IPCC. Simpler models that isolate these processes, and allow us to investigate the fundamental non-linearities, are needed.

Third, we believe that there is much room for conceptual models of convection and clouds, both in terms of capturing uncertainties associated with microphysics and in capturing the essence of moist convection at the base of the hierarchy. It is enticing to think that resolution will 'save' us from convective parameterizations, but it likely that we will find that microphysics is ready to supply uncertainty once our climate models begin to resolve the convective scales. Deliberately simplified convection schemes that can start to take into account uncertainty in microphysics would provide a conceptual test bed, We are also still some distance from a fundamental understanding of such basic matters as convective aggregation, and this may require quite simple models of moist convection to isolate the essential processes.

Finally, as discussed in the introduction, the widespread adoption of a consistent hierarchy in the atmospheric sciences has been limited by the practical issue of sharing models, and keeping them connected with the newest models at the complex end of the hierarchy. It is not simply an issue of documenting and publishing codes, but of enabling other research groups to easily use them on different machines without detailed knowledge of the soft- and hardware. A related issue is allowing our simpler models to upgrade with changes in the modeling framework and underlying numerics. For example, many of the models in the diabatic hierarchy are based on older dynamical cores. The success of the benchmark models highlighted in this review has in large part depended on practical support (e.g., documentation, version control, and code history) which allowed multiple groups to use the same model on their own machines, and perhaps more importantly, to be able to modify the code to facilitate new experiments and create new science.

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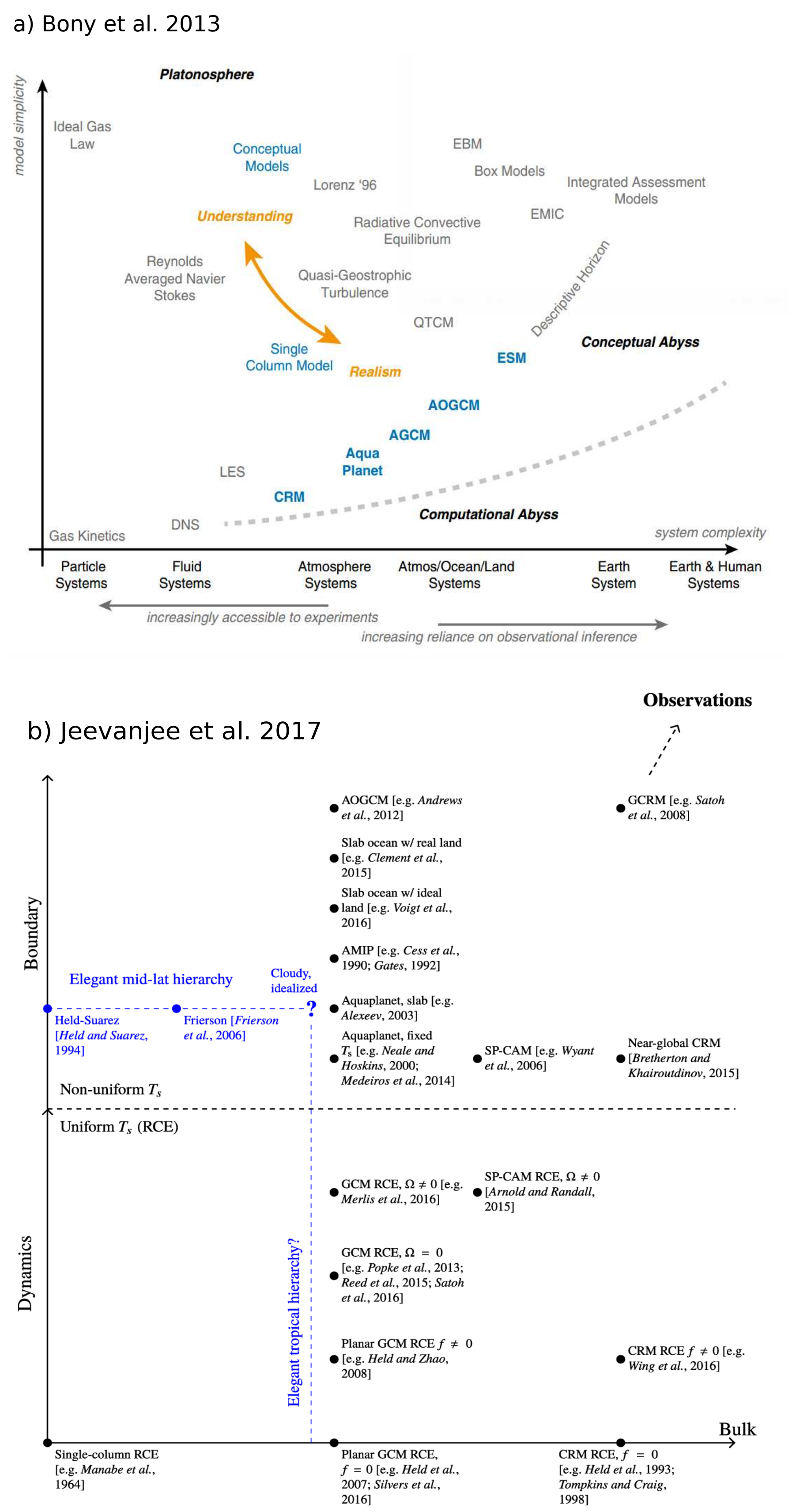


Figure 1. Categorizing atmospheric climate models in terms of complexity a) [*Bony et al.*, 2013] and b) grouped in terms of model configurations for Earth's climate [*Jeevanjee et al.*, 2017].

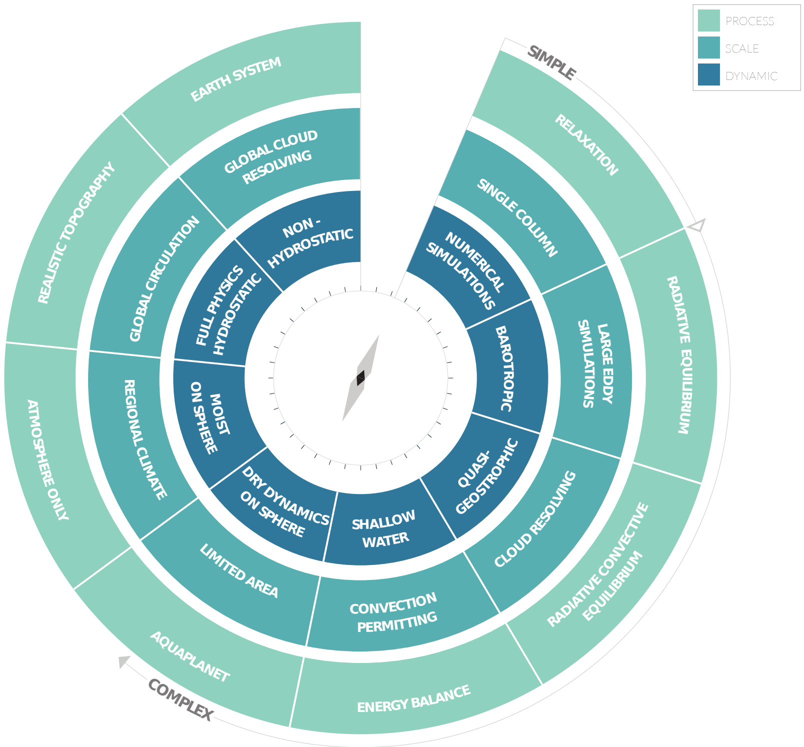


Figure 2. Three principles view of the hierarchies. The outer ring is the process hierarchy, inner ring the hierarchy of scale and central ring the dynamical hierarchy. Clockwise elements show simple configurations that expand to more complicated configurations. This is one possible configuration of each of the hierarchies to illustrate the concept.

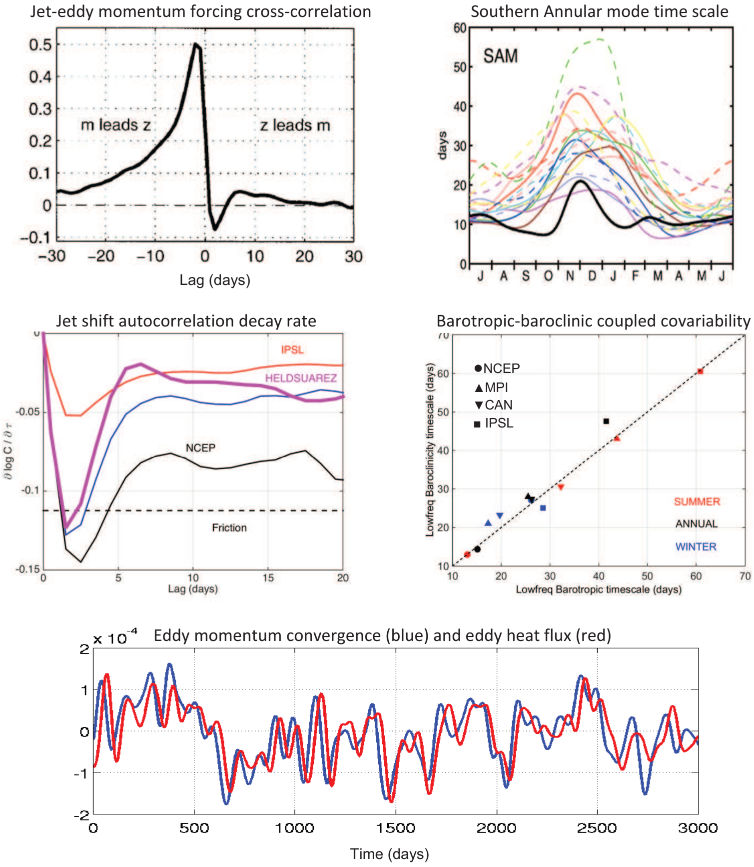


Figure 3. (a) Lagged correlation between zonal mean wind (z) and eddy momentum forcing (m) from *Lorenz and Hartmann* [2001]. (b) Autocorrelation timescale of the Southern Annular Mode for observations (black thick solid) and CMIP3 models (colors). From *Gerber et al.* [2008]. (c) Logarithmic decay rate of autocorrelation for zonal wind anomalies in observations (black), two CMIP5 climate models (IPSL: red, CAN:blue) and the Held and Suarez model (magenta). (d) Scatterplot between low-frequency logarithmic decay rates of baroclinicity and barotropic wind anomalies (average from 5-20 day lags) for the models and seasons indicated. (e) Sample timeseries of the low-frequency eddy momentum (blue) and heat (red) flux contributions to the upper-layer Eliassen-Palm divergence in the QG simulations of *Zurita-Gotor et al.* [2014]

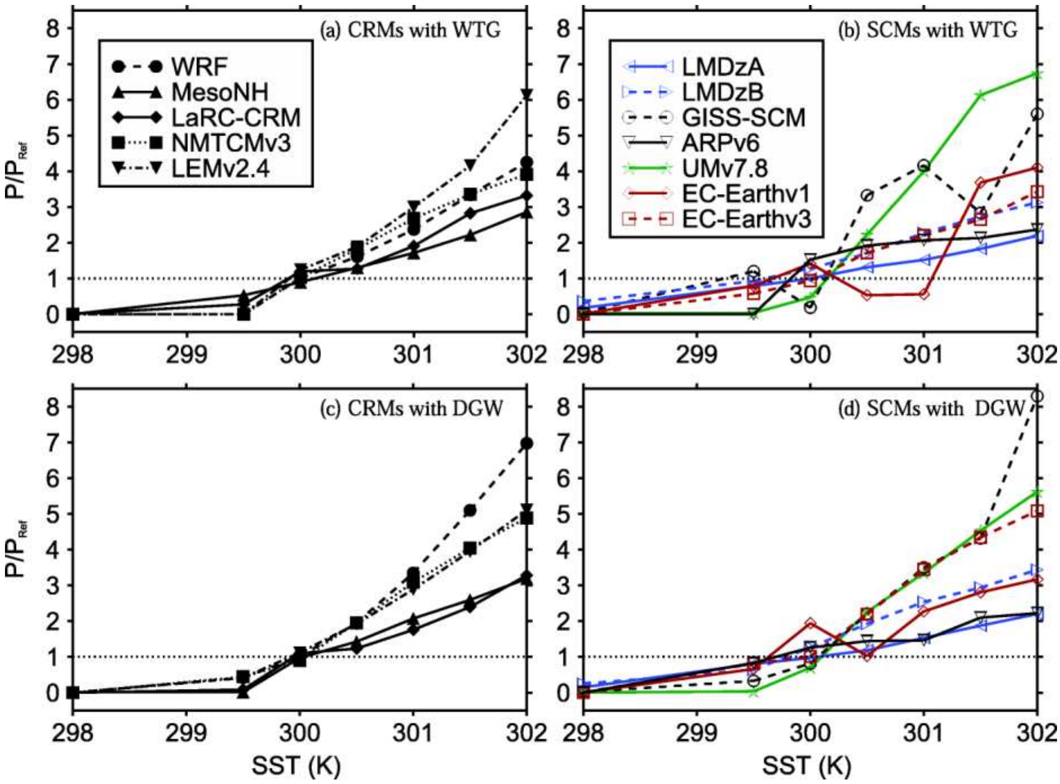
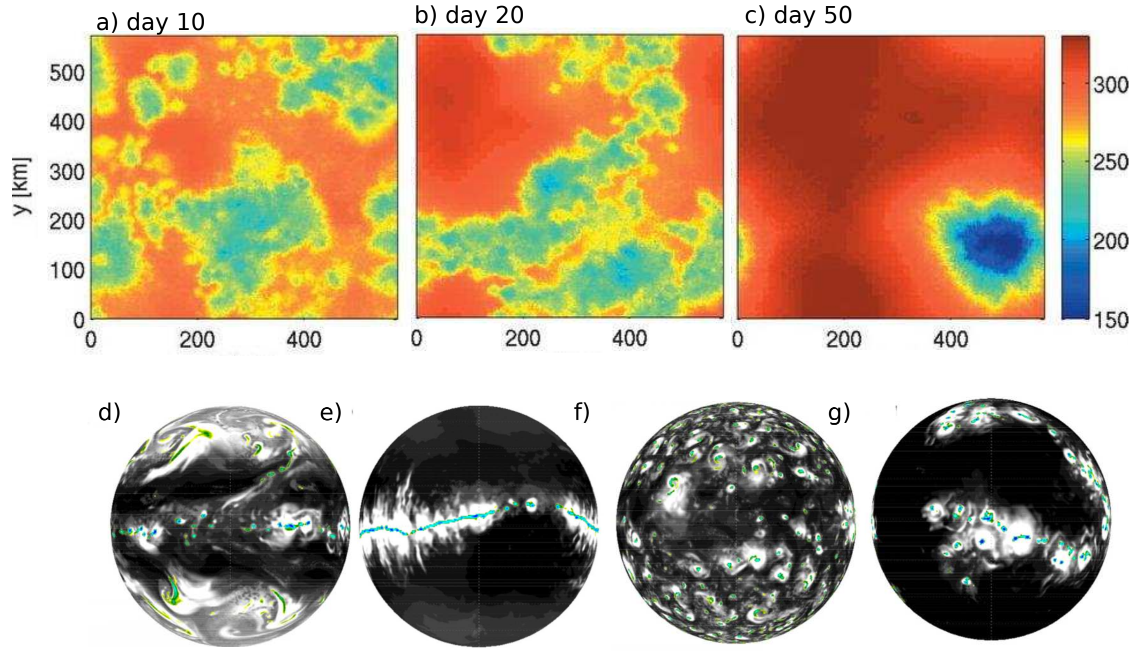


Figure 4. The precipitation rate for simulations using weak temperature gradient (top) and damped gravity wave (bottom) from *Daleu et al.* [2016]



2030 **Figure 5.** Radiative-Convective Equilibrium simulations in a CRM: top panel is daily OLR for a fixed SST
2031 (301K) run after a) 10, b) 20 and c) 50 days of the simulation, adapted from *Bretherton et al.* [2005]. The bottom
2032 panel is OLR for global CRM aquaplanet stimulations using zonally symmetric SSTs similar to observation d)
2033 with rotation (Earth like), e) without rotation, and uniform SSTs f) with and g) without rotation (RCE case),
2034 adapted from *Satoh et al.* [2016].

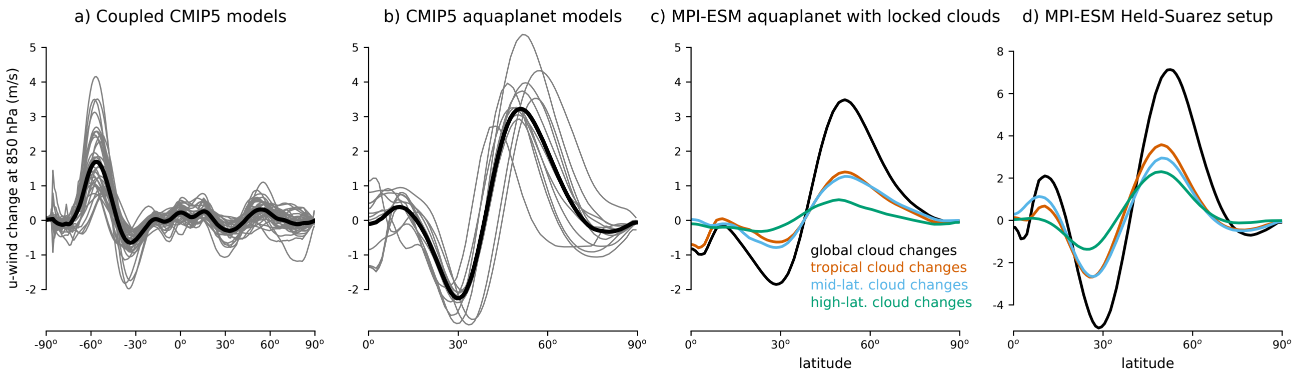


Figure 6. Extratropical cloud-circulation coupling. The impact of clouds on the eddy driven jet stream response to global warming in a hierarchy of GCMs. The zonal-mean time-mean change in 850 hPa zonal wind (ms^{-1}) for each latitude (°) for the ensemble mean (bold line) and individual models (gray) for a) CMIP5 coupled Earth system models with $4\times CO_2$ and b) aquaplanet CMIP5 models with prescribed-SST and 4 K SST warming. For the MPI-ESM model in aquaplanet prescribed-SST setup, simulations with the cloud-locking method and imposed global (black) and regional (colors) cloud changes show the cloud-radiative contribution to the eddy driven jet response to warming (panel c). The global and regional cloud impacts are reproduced in panel d) using a dry Held-Suarez setup of the MPI-ESM model perturbed with the radiative forcing from cloud changes of panel c. Because panels b-d are for aquaplanet simulations, only the Northern hemisphere is shown. Note the different y-scale in panel d, which reflects the increased jet sensitivity of the Held-Suarez setup. Figure adapted from *Voigt and Shaw* [2016].

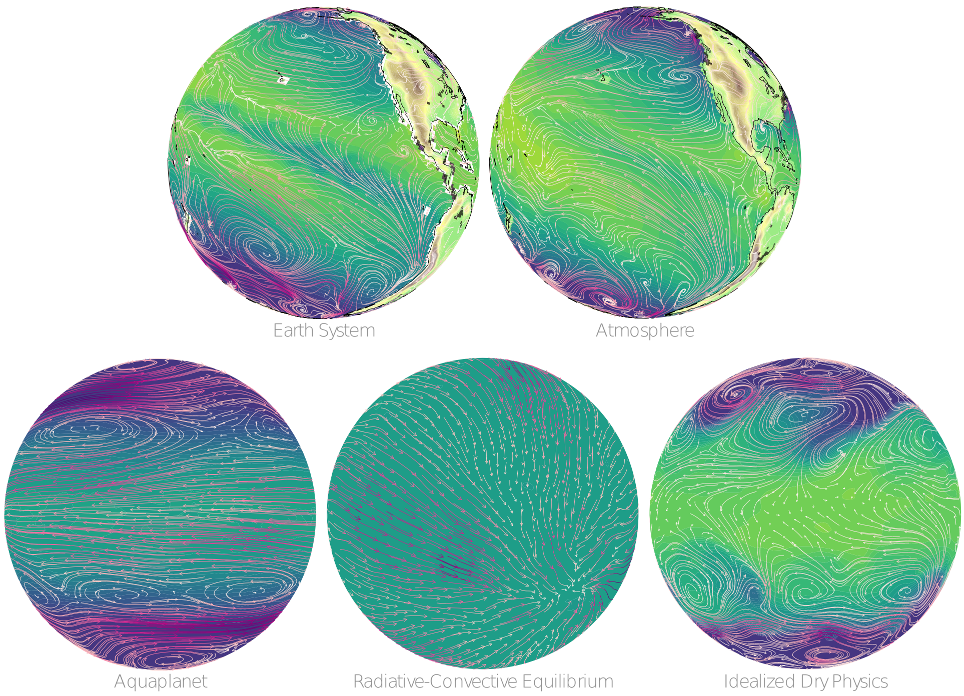


Figure 7. Models available within the hierarchy in the CESM system. (Top) The Earth system model and atmosphere only models (with prescribed SST). (bottom) Aquaplanet, RCE and idealized dry physics. Each globe is a monthly mean except for the idealized dry model which is a snapshot. The colour contours over the ocean are SST and over land topography. Streamlines are the near-surface wind (thicker lines are stronger winds).