

NCAR CESM2.0 release of CAM-SE: A reformulation of the spectral-element dynamical core in dry-mass vertical coordinates with comprehensive treatment of condensates and energy

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Abstract

It is the purpose of this paper to provide a comprehensive documentation of the new NCAR version of the spectral-element (SE) dynamical core as part of the CESM2.0 release. This version differs from previous releases of the SE dynamical core in several ways. Most notably the hybrid-sigma vertical coordinate is based on dry pressure, the condensates are dynamically active in the thermodynamic and momentum equations (also referred to as condensate loading), and the continuous equations of motion conserve a more comprehensive total energy that includes condensates. The code base has been significantly reduced, sped-up, and cleaned up as part of integrating SE as a dynamical core in the CAM (Community Atmosphere Model) repository rather than importing the SE dynamical core from HOMME (high-order method modeling environment) as an external code.

1 Introduction

Over the last two decades or so the spectral-element (SE) method has been considered a candidate next-generation numerical method for the fluid flow solver in global weather/climate models [Fournier *et al.*, 2004; Baer *et al.*, 2006; Kelly and Giraldo, 2012; Giraldo *et al.*, 2013] **other references needed?**. The main motivations were the SE methods near perfect scalability, high-order accuracy for smooth problems, and mesh-refinement capabilities. For a while the Community Earth System Model [CESM; Hurrell *et al.*, 2013] has supported a SE dynamical core option in the atmosphere component CAM [Community Atmosphere Model; Neale *et al.*, 2012] discretized on a cubed-sphere grid (Figure 1a). The SE dynamical core in CAM supports uniform resolution grids based on the equi-angular gnomonic cubed-sphere grid and a mesh-refinement configuration with local increases in resolution through conformal mesh-refinement as shown on Figure 1b [Fournier *et al.*, 2004; Baer *et al.*, 2006; Zarzycki *et al.*, 2013, 2014]. The dynamical core code resided in the High-Order Methods Modeling environment (HOMME), a framework for developing new generation computationally efficient and petascale capable dynamical cores based on the SE method [Thomas and Loft, 2000; Taylor *et al.*, 2008], the discontinuous Galerkin method [Nair *et al.*, 2009] and finite-volume method [Erath *et al.*, 2012; Lauritzen *et al.*, 2017]. Previously, the SE dynamical core was imported into CAM as an external code base. Consequently any updates to the HOMME code base would have to be imported into CAM and pass regression tests in both HOMME and CAM. We refer to this setup as CAM-HOMME, which has been extensively tested in AMIP-style climate simulations [e.g., Evans *et al.*, 2012; Reed *et al.*, 2015; Bacmeister *et al.*, 2016; Rhoades *et al.*, 2016].

With the department of energy (DOE) branching from the CESM to develop an Earth System Model using their own code base, NCAR decided to no longer import HOMME as an external code base in CAM. A reason for doing so was the cumbersome software engineering involved in maintaining two code bases and having to support configurations in HOMME not used in CAM. Instead the SE configuration of HOMME became part of the CAM code base. With that came a massive code clean-up to satisfy software engineering standards in CAM (and make the code more user-friendly for the community) as well as considerable science changes described in this paper.

The SE dynamical core is the default dynamical core for high-resolution CESM applications, in particular, the configuration in which the average distance between grid-points is approximately 25km [Bacmeister *et al.*, 2016]. At such resolutions the effects of condensates (such as cloud liquid and rain) may have a significant effect on the dynamics [Bacmeister *et al.*, 2012]. CAM-HOMME does not represent the effect of condensates in the thermodynamic and momentum equations (also referred to as condensate loading). Including the thermodynamic and mass effect of condensates in the dynamical core is easier when using a dry-mass hybrid-sigma vertical coordinate than when using the usual moist-

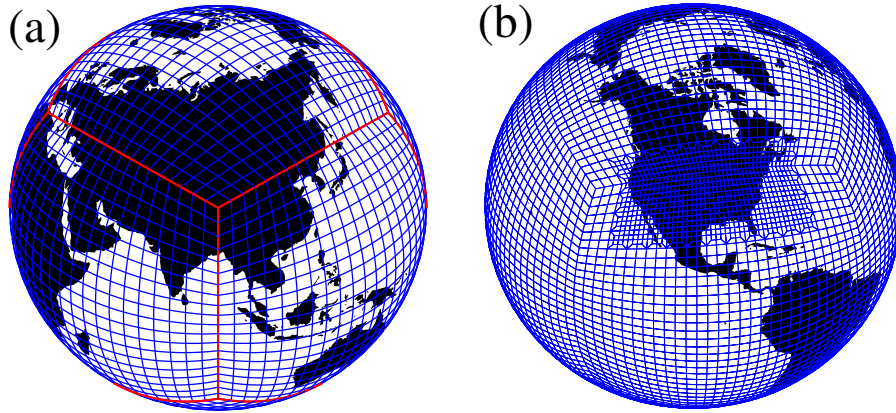


Figure 1. (a) The gnomonic equi-angular cubed-sphere grid that define the elements in CAM-SE. The red lines show the cubed-sphere panel edges. (b) A low resolution conformal mesh-refinement grid (elements) referred to as the CONUS-grid (Contiguous United States) used in CAM-SE.

mass hybrid-sigma vertical coordinate [e.g., *Simmons and Burridge*, 1981]. This is the initial motivation for using a dry-mass vertical coordinate.

A second motivation for switching to a dry-mass vertical coordinate is the consistent coupling with the physical parameterizations. CAM physics assumes that the moist pressure levels are constant during the parameterization updates. Consequently the moist pressure levels stay constant even when moisture leaves the column (e.g., rains out). At the very end of CAM physics the change in water vapor in each column is taken into account by scaling the mixing ratios for all tracers that are based on specific/moist mixing ratios so that dry air mass and tracer mass is conserved [see Section 3.1.6 in *Neale et al.*, 2012]. This scaling does not guarantee shape-preservation and changes the total energy. In addition, the pressure field in CAM physics does not take into account the mass of condensates. When using a dry-mass vertical coordinate, the coordinate surfaces (assuming dry mass is constant) remain constant throughout the physics updates and there is no need to adjust tracer mixing ratios, and one can more rigorously take into account the work performed by water variables in the context of the energy cycle.

The third motivation for using a dry-mass vertical coordinate relates to total energy conservation. Currently the energy fixer in CAM is based on a dry total energy [*Williamson et al.*, 2015] that uses the same heat-capacity for dry air and water vapor, and does not include the effect of condensates. To move towards a more accurate treatment of energy in CAM, a first step is to develop a dynamical core based on equations of motion conserving an energy that more accurately represents water vapor as well as condensates. This is most easily done when using a dry-mass vertical coordinate so that the energies associated with all water variables are clearly separated.

In this paper we present a version of the SE dynamical core using a dry-mass vertical coordinate that includes condensate-loading, and the continuous equations of motion conserve a comprehensive moist total energy containing all prognostic water variables and their respective heat capacities in the thermodynamic equation. As this paper serves as documentation for the CESM2.0 version of SE we also provide details on the spectral-element method and viscosity operators that have not been comprehensively documented in previous publications.

The paper is organized as follows. In section 2 the continuous equations of motion are derived which involves a detailed discussion of moist thermodynamics in the pres-

ence of condensates. The axial angular momentum (AAM) and total energy conservation properties of continuous system of equations is also discussed. In section 3 the discretized equations of motion with focus on the vertical discretization in dry-mass vertical coordinates are derived. Details on the horizontal SE discretization on the cubed-sphere are also presented. Section 4 presents some results, validating the new dynamical core in idealized configurations. First, a moist baroclinic wave with simple warm-rain micro-physics and secondly in a CAM6¹ aqua-planet configuration. The computational performance of CAM-SE is presented in section 4. The paper ends with summary and conclusions in section 5.

2 Continuous equations

Before writing the continuous equations of motion using a dry-mass vertical coordinate, we first need to discuss the representation of water variables (section 2.1), discuss the ideal gas law and derive the thermodynamic equation for a mixture containing all water species (section 2.2). The discussion of the equations of motion in the presence of water vapor, cloud liquid, ice, rain and snow closely follows *Staniforth et al.* [2006]. Thereafter the dry mass vertical coordinate is defined in section 2.4. Details on viscosity, frictional heating and global conservation properties are presented in this section as well.

2.1 Representation of water phases in terms of dry and wet (specific) mixing ratios

Define the dry mixing ratios for the water variables (vapor 'wv', cloud liquid 'cl', cloud ice 'ci', rain 'rn' and snow 'sw')

$$m^{(\ell)} \equiv \frac{\rho^{(\ell)}}{\rho^{(d)}}, \text{ where } \ell = \text{'wv', 'cl', 'ci', 'rn', 'sw'}, \quad (1)$$

where $\rho^{(d)}$ is the mass of dry air per unit volume of moist air and $\rho^{(\ell)}$ is the mass of the water substance of type ℓ per unit volume of moist air. Note that the mixing ratio for dry air is one: $m^{(d)} \equiv \frac{\rho^{(d)}}{\rho^{(d)}} = 1$. Moist air refers to air containing dry air, water vapor, cloud liquid, cloud ice, rain amount and snow amount. For notational purposes define the set of all components of air

$$\mathcal{L}_{all} = \{\text{'d', 'wv', 'cl', 'ci', 'rn', 'sw'}\}, \quad (2)$$

a set only referring to all water variables,

$$\mathcal{L}_{water} = \{\text{'wv', 'cl', 'ci', 'rn', 'sw'}\}, \quad (3)$$

and a set referring to all condensates (non-gas components of water)

$$\mathcal{L}_{cond} = \{\text{'cl', 'ci', 'rn', 'sw'}\}. \quad (4)$$

The density of a unit volume of moist air is related to the dry air density through

$$\rho = \rho^{(d)} \left(\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} \right). \quad (5)$$

Mixing ratio's can also be specified in terms of density per density of moist air, in other words, specific/moist mixing ratios

$$q^{(\ell)} \equiv \frac{\rho^{(\ell)}}{\rho}, \quad (6)$$

¹ the CAM6 physics tag is cam5_4_128 which is a near final pre-release version of the CESM2.0-CAM6 physics package; this paper focuses on the dynamical core and the details of the physics package are less important

where, in particular, $q^{(wv)}$ is the specific humidity.

It is straight forward to convert between moist and dry mixing ratios

$$m^{(\ell)} = \frac{q^{(\ell)}}{1 - \sum_{\ell \in \mathcal{L}_{water}} q^{(\ell)}}, \quad (7)$$

$$q^{(\ell)} = \frac{m^{(\ell)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}}. \quad (8)$$

Note that if water vapor undergoes a phase change to rain and leaves the column, then the specific/wet mixing ratios change but the dry mixing ratios do not.

2.2 Ideal gas law and virtual temperature

In this section the ideal gas law is derived for moist air containing condensates. Part of that derivation is the definition of virtual temperature. If we assume that moist air obeys Amagat's law (or the law of partial volumes) then a unit volume of moist air $V = 1$ is the sum the volume of dry air and the volumes of the different forms of water in the air

$$V = \sum_{\ell \in \mathcal{L}_{all}} V^{(\ell)}. \quad (9)$$

Now let $\alpha \equiv \frac{1}{\rho}$ denote the specific volume of moist air, let $\alpha^{(gas)}$ be the specific volume of the gaseous components of moist air (the volume of water vapor and dry air per unit mass of water vapor and dry air), let $\alpha^{(cl)}$ be the specific volume of cloud liquid (i.e. volume occupied by unit mass of cloud liquid), and similarly for cloud ice, rain and snow. Then we can write (9) in terms of densities and specific volumes

$$\rho\alpha = \rho^{(gas)}\alpha^{(gas)} + \sum_{\ell \in \mathcal{L}_{cond}} \rho^{(\ell)}\alpha^{(\ell)}, \quad (10)$$

where $\rho^{(gas)} = \rho^{(d)} + \rho^{(wv)}$. Note that $\alpha^{(\ell)} \neq \frac{1}{\rho^{(\ell)}}$ since $\rho^{(\ell)}$ is defined in terms mass of ℓ per unit volume of moist air and not mass of ℓ per unit volume of water phase ℓ . Dividing (10) with $\rho^{(d)}$ and substituting (1) yields

$$\left(\sum_{\ell} m^{(\ell)} \right) \alpha = (1 + m^{(wv)}) \alpha^{(gas)} + \sum_{\ell \in \mathcal{L}_{cond}} m^{(\ell)} \alpha^{(\ell)}. \quad (11)$$

The ideal gas law for the gaseous components, using Dalton's law of partial pressures, is

$$p = p^{(d)} + p^{(wv)} = \rho^{(d)} R^{(d)} T + \rho^{(wv)} R^{(wv)} T, \quad (12)$$

where $p^{(d)}$ is the dry pressure and $p^{(wv)}$ is the water vapor pressure. We may write (12) in terms of $\rho^{(gas)}$ by multiplying (12) with $\frac{\rho^{(gas)}}{\rho^{(gas)}}$ and write the equation in terms of mixing ratios

$$p = \rho R^{(d)} T \rho^{(gas)} \left(\frac{\rho^{(d)}}{\rho^{(gas)}} + \frac{\rho^{(wv)}}{\rho^{(gas)}} \frac{R^{(wv)}}{R^{(d)}} \right) = \frac{R^{(d)} \rho^{(gas)} T \left(1 + \frac{1}{\epsilon} m^{(wv)} \right)}{(1 + m^{(wv)})}. \quad (13)$$

where $\epsilon \equiv \frac{R^{(d)}}{R^{(wv)}}$. Since (12) holds for the gaseous components only we may substitute $\frac{1}{\rho^{(gas)}}$ with $\alpha^{(gas)}$ to get:

$$p = \frac{R^{(d)} T \left(1 + \frac{1}{\epsilon} m^{(wv)} \right)}{(1 + m^{(wv)}) \alpha^{(gas)}}. \quad (14)$$

Using (11) to eliminate $(1 + m^{(wv)}) \alpha^{(gas)}$ in (14) and simplifying yields

$$p = \rho R^{(d)} T_v, \quad (15)$$

where the virtual temperature is given by

$$T_v = T \left[\frac{1 + \frac{1}{\epsilon} m^{(wv)}}{1 + m^{(wv)} + \sum_{\ell \in \mathcal{L}_{cond}} m^{(\ell)} \left(1 - \frac{\alpha^{(\ell)}}{\alpha}\right)} \right]. \quad (16)$$

The ratio of the density of moist air to the density of the condensates are negligible (e.g., the terms $\frac{\alpha^{(\ell)}}{\alpha}$ are of order 10^{-3} or less) hence the formula for virtual temperature is simplified to

$$T_v = T \left(\frac{1 + \frac{1}{\epsilon} m^{(wv)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} \right). \quad (17)$$

2.3 Thermodynamic equation

The first law of thermodynamics for a mixture of dry air, water vapor, cloud liquid water, cloud ice, rain and snow is given by

$$\frac{\sum_{\ell \in \mathcal{L}_{all}} c_v^{(\ell)} m^{(\ell)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} \delta T + p \delta \alpha = \delta Q, \quad (18)$$

where δQ is the amount of heat per unit mass that is supplied reversibly to the moist air, δT and $\delta \alpha$ are the moist air temperature and specific volume changes, and $c_v^{(\ell)}$ is the heat capacity at constant volume for component ℓ of moist air. Note that

$$\left(\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} \right) p \delta \alpha = p \delta \left[\left(\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} \right) \alpha \right], \quad (19)$$

$$= (1 + m^{(wv)}) p \delta \alpha^{(gas)}, \quad (20)$$

$$= (R^{(d)} + m^{(wv)} R^{(wv)}) \left(\delta T - \frac{T}{p} \delta p \right). \quad (21)$$

where, in (19), we have assumed that $m^{(\ell)}$ is constant. To get (20) we substituted (11) into the right-hand side of (19) and assumed that the condensates are incompressible ($\delta \alpha^{(\ell)} = 0$ for $\ell \in \mathcal{L}_{cond}$). Finally (14) has been used to substitute for $\alpha^{(gas)}$ in (21). Using (21) in the first law of thermodynamics (18), substituting $R^{(d)} = c_v^{(d)} - c_v^{(wv)}$ and $R^{(wv)} = c_p^{(wv)} - c_v^{(wv)}$, using that $c_p^{(\ell)} = c_v^{(\ell)}$ for the condensates (they are incompressible) and rearranging terms yields

$$\delta T - \frac{(R^{(d)} + m^{(wv)} R^{(wv)}) T}{\left(\sum c_p^{(\ell)} m^{(\ell)} \right) p} \delta p = \frac{\left(\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} \right)}{\left(\sum_{\ell \in \mathcal{L}_{all}} c_p^{(\ell)} m^{(\ell)} \right)} \frac{\delta Q}{T}. \quad (22)$$

Equation (22) can be written in a more familiar form

$$\delta T - \frac{R}{c_p} \frac{T}{p} \delta p = \frac{1}{c_p T} \delta Q, \quad (23)$$

if we define c_p and R as

$$c_p = \frac{\sum_{\ell \in \mathcal{L}_{all}} c_p^{(\ell)} m^{(\ell)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}}, \quad (24)$$

$$R = \frac{\sum_{\ell \in \mathcal{L}_{all}} R^{(\ell)} m^{(\ell)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}}. \quad (25)$$

where $R^{(\ell)} = 0$ for $\ell \in \mathcal{L}_{cond}$. Note that with these definitions of c_p and R the ideal gas law can be written as

$$p = \rho R T, \quad (26)$$

and

$$c_p = R + c_v. \quad (27)$$

The thermodynamic equation (23) may also be written in terms of full density

$$\delta T - \frac{1}{\rho c_p} \delta p = \frac{1}{c_p T} \delta Q. \quad (28)$$

2.4 Vertical coordinate

2.4.1 Definition

Let $\mathcal{P}_s^{(d)}$ be the surface pressure of a dry atmosphere and $\mathcal{P}_t^{(d)}$ the pressure at the model top. We assume that there is no moisture or condensate above the model top so $p_t^{(d)} = p_t^{(wv)} = \mathcal{P}_t^{(d)}$. Consider a general, terrain-following, vertical coordinate $\eta^{(d)}$ that is a function of dry atmosphere pressure $\mathcal{P}^{(d)}$

$$\eta^{(d)} = h(\mathcal{P}^{(d)}, \mathcal{P}_s^{(d)}), \quad (29)$$

where $h(\mathcal{P}_s^{(d)}, \mathcal{P}_s^{(d)}) = 1$ and $h(\mathcal{P}_t^{(d)}, \mathcal{P}_s^{(d)}) = 0$. Note that by removing the superscript (d) from the equations above (so that the dry variable represent moist pressure variables) then the vertical coordinate is the usual hybrid-pressure coordinate widely used in hydrostatic global modeling [Simmons and Burridge, 1981]. The top and bottom boundary conditions are that $\eta(\mathcal{P}_s^{(d)}, \mathcal{P}_s^{(d)}) = 0$ and $\eta(\mathcal{P}_t^{(d)}, \mathcal{P}_s^{(d)}) = 0$. Note that using a dry-mass vertical coordinate simplifies the coupling to physics since the dry mass coordinate remains constant even if water leaves the column.

2.4.2 Dry pressure and dry atmosphere pressure

The observant reader will have noticed that we denote the dry atmosphere pressure $\mathcal{P}^{(d)}$ and not $p^{(d)}$. The moist pressure at given height z' can be computed from the hydrostatic balance

$$p = \int_{z'=z}^{z'=\infty} \rho g dz', \quad (30)$$

$$= \int_{z'=z}^{z'=\infty} \rho_d \left(\sum_{\ell \in \mathcal{L}_{all}} m_\ell \right) g dz', \quad (31)$$

$$= \sum_{\ell \in \mathcal{L}_{all}} \mathcal{P}^{(\ell)}, \quad (32)$$

where the right-hand side of (32) is the weight of dry air, water vapor, cloud liquid, cloud ice, rain and snow per unit area, respectively:

$$\mathcal{P}^{(\ell)} = \int_{z'=z}^{z'=\infty} \rho_d m_\ell g dz'. \quad (33)$$

Using Dalton's law of partial pressures, (32) can be written as

$$p^{(d)} + p^{(wv)} = \sum_{\ell \in \mathcal{L}_{all}} \mathcal{P}^{(\ell)}. \quad (34)$$

From (34) it is clear that in the presence of condensate one can not equate $p^{(d)}$ with $\mathcal{P}^{(d)}$, nor $p^{(wv)}$ with $\mathcal{P}^{(wv)}$. The dry pressure and water vapor pressure are both affected by the weight of the condensates even though the condensates do not exert a pressure. Hence the dry pressure $p^{(d)}$ is different from the pressure in a dry atmosphere $\mathcal{P}^{(d)}$. The hydrostatic balance of a dry atmosphere, written in terms of differentials, is given by

$$d\mathcal{P}^{(d)} = -\rho_d g dz, \quad (35)$$

whereas, in a moist atmosphere, a dry pressure hydrostatic equation does not hold

$$dp^{(d)} \neq -\rho_d g dz. \quad (36)$$

The differential of the moist pressure can be written in terms of the dry atmosphere pressure though:

$$dp = -\rho g dz = -\rho_d \left(\sum_{\ell \in \mathcal{L}_{all}} m_\ell \right) g dz = d\mathcal{P}^{(d)} \left(\sum_{\ell \in \mathcal{L}_{all}} m_\ell \right). \quad (37)$$

2.5 Equations of motion

The $\eta^{(d)}$ -coordinate adiabatic and frictionless atmospheric primitive equations assuming floating Lagrangian vertical coordinates [Starr, 1945; Lin, 2004] can be written in vector invariant form as

$$\frac{\partial \vec{v}}{\partial t} + (\zeta + f) \hat{k} \times \vec{v} + \nabla_{\eta^{(d)}} \left(\frac{1}{2} \vec{v}^2 + \Phi \right) + \frac{1}{\rho} \nabla_{\eta^{(d)}} p = 0, \quad (38)$$

$$\frac{\partial T}{\partial t} + \vec{v} \cdot \nabla_{\eta^{(d)}} T - \frac{1}{c_p \rho} \omega = 0, \quad (39)$$

$$\frac{\partial}{\partial t} \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} m^{(\ell)} \right) + \nabla_{\eta^{(d)}} \cdot \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} m^{(\ell)} \vec{v} \right) = 0, \quad \ell \in \mathcal{L}_{all}, \quad (40)$$

where Φ is the geopotential height ($\Phi = g z$, where g is the gravitational constant), \hat{k} is the unit vector normal to the surface of the sphere, $\vec{v} = (u, v)$ is the velocity vector with u being the zonal velocity component and v the meridional velocity component, $\zeta = \hat{k} \cdot \nabla \times \vec{v}$ is vorticity, f Coriolis parameter, and $\omega = dp/Dt$ is the pressure vertical velocity with $d/dt = \frac{\partial}{\partial t} + \vec{v} \cdot \nabla$ being the material/total derivative.

The prognostic equations for \vec{v} , the temperature T , pseudo density of a dry atmosphere $\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}}$, and pseudo tracer density $\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} m^{(\ell)}$ are solved with the diagnostic equation for geopotential height (hydrostatic balance)

$$\frac{\partial \Phi}{\partial \eta^{(d)}} = - \frac{R^{(d)} T_v}{p} \frac{\partial p}{\partial \eta^{(d)}}, \quad (41)$$

where

$$\frac{\partial p}{\partial \eta^{(d)}} = \frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \left(\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} \right). \quad (42)$$

For diagnosing vertical pressure velocity ω we note that

$$\omega(\eta^{(d)}) = \frac{dp}{dt}(\eta^{(d)}), \quad (43)$$

$$= \int_{\eta^{(d)}}^{\eta^{(d)=0}} \frac{d}{dt} \left(\frac{\partial p}{\partial \eta^{(d)}} \right) d\eta^{(d)}, \quad (44)$$

$$= \int_{\eta^{(d)}}^{\eta^{(d)=0}} \frac{\partial}{\partial t} \left(\frac{\partial p}{\partial \eta^{(d)}} \right) d\eta^{(d)} + \int_{\eta^{(d)}}^{\eta^{(d)=0}} \vec{v} \cdot \nabla \left(\frac{\partial p}{\partial \eta^{(d)}} \right) d\eta^{(d)}. \quad (45)$$

2.6 Hyperviscosity and frictional heating

The spectral-element method does not have implicit diffusion. Hyperviscosity operators are applied to the prognostic variables to dissipate energy near the grid scale. Hyperviscosity also damp the propagation of spurious grid-scale modes [Ainsworth and Wajid, 2009] and smooth the solution at element boundaries where the basis-functions are least smooth (C^0 -continuous). For the uniform resolution configuration constant hyperviscosity coefficients are used whereas the variable resolution configuration uses either a scaling of the coefficients according to a local element length scale [Zarzycki et al., 2013] or a tensor-based hyperviscosity operator approach [Guba et al., 2014a].

2.6.1 Viscosity operator

On the right-hand side of the equations of motion hyperviscosity terms are added. For the momentum equations (38) the viscous terms are as follows. By using the vector identity $\nabla^2 \vec{v} = \nabla(\nabla \cdot \vec{v}) - \nabla \times (\nabla \times \vec{v})$, the viscosity term splits into two terms. The first term damps divergent modes and the latter term damps rotational modes. Thereby one can

damp divergent modes more (or less) than rotational modes by having different coefficients (ν_d and ν_v , respectively) in front of the respective terms

$$\nu_d \nabla(\nabla \cdot \vec{v}) - \nu_v \nabla \times (\nabla \times \vec{v}). \quad (46)$$

The 4th-order hyperviscosity operator is computed by iteratively applying the Laplacian operator (46).

The pseudo-density is damped with $\nu_p \nabla^4 \left\{ \frac{\partial p}{\partial \eta^{(d)}} \right\}$ and temperature with $\nu_T \nabla^4 T$. The specific damping coefficients for divergence (ν_{div}), vorticity (ν_{vor}), pressure-level thickness (ν_p) and temperature (ν_T) are resolution specific and provided in Appendix A.3.

The horizontal hyperviscosity operator can be applied on η_d -surfaces, $\nabla^4 = \nabla_{\eta_d}^4$, but it may be advantageous to apply the hyperviscosity operator on approximate pressure surfaces

$$\nu \nabla^4 \Xi = \nu \nabla_{\eta_d}^4 \Xi - \frac{\partial \Xi}{\partial p} \nu \nabla_{\eta_d}^4 p, \quad \Xi = \vec{v}, T, \quad (47)$$

[p.58 in *Neale et al.*, 2012] to reduce spurious diffusion over steep topography. In theory the damping of the pseudo-density should be zero if hyperviscosity is applied on pressure surfaces. However, for stability it is necessary to damp pressure-level thicknesses but instead of applying ∇^4 to $\frac{\partial p}{\partial \eta^{(d)}}$ it is applied to the difference between $\frac{\partial p}{\partial \eta^{(d)}}$ and a smoothed version of $\frac{\partial p}{\partial \eta^{(d)}}$ referred to as $\left(\frac{\partial p}{\partial \eta^{(d)}} \right)^{(ref)}$. The reference/smoothed pressure-level thickness is defined in Appendix A.2.

In the top three layers second-order diffusion (Laplacian operator) is applied to the prognostic variables to provide a sponge layer. The sponge layer plays an important role in controlling the polar night jet in low top models [see, e.g., *Lauritzen et al.*, 2011].

2.6.2 Frictional heating

Let $\delta \vec{v}$ be the change in the velocity vector due to diffusion of momentum. Then the change in kinetic energy due to hyperviscosity applied to \vec{v} is $\frac{1}{2} \rho \vec{v} \cdot \delta \vec{v}$. This kinetic energy is converted to a heating rate by adding a heating term $\delta \mathcal{T}$ in the thermodynamic equation corresponding to the kinetic energy change

$$\rho c_p \delta \mathcal{T} = -\frac{1}{2} \rho \vec{v} \cdot \delta \vec{v} \Rightarrow \delta \mathcal{T} = -\frac{1}{c_p} (\vec{v} \cdot \delta \vec{v}), \quad (48)$$

[p.71 in *Neale et al.*, 2012]. As shown in the results section this term is rather large and therefore important for good energy conservation characteristics of the dynamical core.

Before discretizing the equations of motion some important conservation properties of the equations are discussed. For models intended for long simulations good energy conservation properties are considered beneficial so that global energy fixers do less non-local restoration of global energy [e.g. *Thuburn*, 2008]. As an aside it is noted that the global energy fixer take up a significant fraction of runtime at ultra-high resolution as the operation requires a global communication. Good axial angular-momentum conservation may be important for the simulation of the Quasi-Biennial Oscillation (QBO) although the accurate simulation of the QBO also depends on vertical resolution, location of model top, and parameterizations (such as nonorographic gravity wave drag). For the simulation of superrotating atmospheres (e.g., Venus), however, the conservation of angular momentum is crucial [*Lebonnois et al.*, 2012]. Below we discuss the conservation properties of the moist equations of motion used in this paper.

2.7 Global conservation: AAM and energy

The conservation law for AAM

$$\mathcal{M} = (u + \Omega r \cos \varphi) r \cos \varphi, \quad (49)$$

(where r is the mean radius of Earth, Ω angular velocity and φ latitude), integrated over the entire domain is derived in Appendix B: and the final equation is

$$\frac{\partial}{\partial t} \int_{\eta=0}^{\eta=1} \iint_S \left[\left(\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} \right) \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) \mathcal{M} \right] dA d\eta^{(d)} = - \iint_S \left[p_s \frac{\partial z_s}{\partial \lambda} \right] dA, \quad (50)$$

where S is the global domain and $dA = r^2 \cos \varphi d\lambda d\varphi$ is an infinitesimal spherical area. The equation clearly separates the AAM of each component of moist air. The term in the square brackets on the right-hand side of (50) is referred to as the mountain torque. In the absence of topography, $z_s = 0m$, (50) states that the angular momentum integrated over the entire domain is constant for the continuous equations of motion. Note that the AAM can be separated into a part (\mathcal{M}_r) associated with the motion of the atmosphere relative to Earth's surface (also known as wind AAM) and another part (\mathcal{M}_Ω) associated with the angular velocity Ω of Earth's surface (referred to as mass AAM)

$$\mathcal{M} = \mathcal{M}_r + \mathcal{M}_\Omega = (ur \cos \varphi) + (\Omega r^2 \cos^2 \varphi). \quad (51)$$

The total energy equation integrated over the global domain is also derived in Appendix B: . The final equation is

$$\frac{1}{g} \frac{\partial}{\partial t} \int_{\eta=0}^{\eta=1} \iint_S \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) \sum_{\ell \in \mathcal{L}_{all}} \left[m^{(\ell)} (K + c_p^{(\ell)} T + \Phi_s) \right] dA d\eta^{(d)} = 0. \quad (52)$$

Note that the energy terms (inside square brackets) in (52) separate into contributions from each component of moist air

$$\left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) \sum_{\ell \in \mathcal{L}_{all}} \left[m^{(\ell)} (K + c_p^{(\ell)} T + \Phi_s) \right]. \quad (53)$$

The total energy equation (52) shows that the equations of motion used in this paper conserve a moist total energy that includes condensates. That said, the CAM physics package energy fixer assumes that the ‘perfect’ adiabatic dynamical core conserves an energy where $c_p^{(wv)} = c_p^{(d)}$, $c_p^{(\ell)} = 0$ for $\ell \in \mathcal{L}_{cond}$ and $m^{(\ell)} = 0$ for $\ell \in \mathcal{L}_{cond}$ in (52) [Williamson *et al.*, 2015]. In other words, the CAM energy fixer uses a dry total energy where the heat capacity for dry air is used for water vapor. The discrepancy between the more comprehensive energy formula (52) and the CAM physics formula for total energy is about 0.5 W/m^2 [Taylor, 2011]. By only including dry air and moisture in ρ and setting $c_p^{(wv)} = c_p^{(d)}$ in the equations of motion, the dynamical core (in the absence of truncation errors) will conserve the energy used in CAM physics. In the model code this is controlled with the logical parameter `lcp_moist`.

3 Discretized equations of motion

3.1 Vertical discretization

In the vertical the atmosphere is discretized into $nlev$ floating Lagrangian layers. The vertical index is 1 for the upper most level and $nlev$ in the lower most level. The level interfaces are referred to as half-levels so that layer k is bounded by interface level $k + 1/2$ and $k - 1/2$. Since we are using a dry-mass vertical reference coordinate, the dry atmosphere reference pressure at the layer interfaces is defined in terms of the hybrid coefficients A and B that are only a function of level index

$$\mathcal{P}_{k+1/2}^{(d)} = A_{k+1/2} p_t + B_{k+1/2} \mathcal{P}_s^{(d)}, \quad (54)$$

and similarly for full levels k . Note that if the ‘ d ’ is removed from the above equation so that the levels are based on moist pressure then the vertical coordinate becomes the usual

hybrid vertical coordinate used in many global hydrostatic models. Every `se_rsplitt` dynamics time-steps the prognostic variables on the floating Lagrangian levels are remapped to the dry-mass reference coordinate (54) (see Section 3.1.5).

CAM-SE is based on a Lorenz vertical staggering – the full level prognostic variables are \vec{v}_k (velocity vector), T_k (temperature), $\Delta\mathcal{P}_k^{(d)} = \mathcal{P}_{k+1/2}^{(d)} - \mathcal{P}_{k-1/2}^{(d)}$ (dry atmosphere pressure level thickness), and $\Delta\mathcal{P}_k^{(d)} m_k^{(\ell)}$ (tracer mass per unit area). At the level interfaces, the geopotential, pressure, vertical velocity and density are defined. In the equations of motion, the full level pressure, geopotential height, vertical velocity and density are needed and choices must be made on how these variables are derived (in discretized space) from the prognostic variables. Here we use the energy and angular momentum conserving method of *Simmons and Burridge* [1981] to compute these quantities at full levels. For simplicity, we do not consider the discretization in the horizontal in this section.

3.1.1 Pressure

The half-level moist pressure is

$$p_{k+1/2} = p_t + \sum_{j=1}^{nlev} \Delta\mathcal{P}_j^{(d)} \left(\sum_{\ell \in \mathcal{L}_{all}} m_j^{(\ell)} \right), \quad (55)$$

where p_t is the pressure at the model top, $\eta^{(d)} = 0$. The full level moist pressure is obtained by averaging [Simmons and Burridge, 1981]

$$p_k = \frac{p_{k+1/2} + p_{k-1/2}}{2}. \quad (56)$$

3.1.2 Geopotential height

Discretizing (41) in the vertical yields

$$\Phi_{k+1/2} = \Phi_s + R^{(d)} \sum_j \left(\frac{T_k}{p_k} \right) \Delta p_j, \quad (57)$$

where the full pressure p_k is given in (56). The half level geopotential is computed by averaging

$$\Phi_k = \frac{\Phi_{k+1/2} + \Phi_{k-1/2}}{2}. \quad (58)$$

3.1.3 Vertical pressure velocity

The vertical pressure velocity ω is obtained by discretizing (45). The first term on the right-hand side of (45) can be computed by using the continuity equations for dry pressure level thickness and water masses in each layer (40)

$$\frac{\partial}{\partial t} \left[\sum_{j=1}^k \Delta\mathcal{P}_j^{(d)} \left(\sum_{\ell \in \mathcal{L}_{all}} m_j^{(\ell)} \right) \right] = - \sum_{j=1}^k \nabla_{\eta^{(d)}} \cdot \left[\Delta\mathcal{P}_j^{(d)} \left(\sum_{\ell \in \mathcal{L}_{all}} m_j^{(\ell)} \right) \right], \quad (59)$$

so that the vertical pressure velocity at half-levels is given by

$$\omega_{k+1/2} = - \sum_{j=1}^k \nabla_{\eta^{(d)}} \cdot \left[\Delta\mathcal{P}_j^{(d)} \left(\sum_{\ell \in \mathcal{L}_{all}} m_j^{(\ell)} \right) \right] + \sum_{j=1}^k \vec{v}_k \cdot \nabla_{\eta^{(d)}} \left[\Delta\mathcal{P}_j^{(d)} \left(\sum_{\ell \in \mathcal{L}_{all}} m_j^{(\ell)} \right) \right], \quad (60)$$

and full level ω is

$$\omega_k = \frac{\omega_{k+1/2} + \omega_{k-1/2}}{2}. \quad (61)$$

3.1.4 Density

Full level density is computed from the ideal gas law (15)

$$\rho_k = \frac{p_k}{R^{(d)} T_k^{(v)}}, \quad (62)$$

where the full pressure p_k is computed as described in section 3.1.1. The virtual temperature is based on prognostic variables defined at the layer centers, so simple substitution into (17) yields $T_k^{(v)}$, and similarly for the computation of $(c_p)_k$.

3.1.5 Vertical remapping

To avoid excessive deformation or even crossing of the floating Lagrangian levels, the prognostic variables defined at

$$\mathcal{P}_{k+1/2}^{(d)} = p_t + \sum_{j=1}^k \Delta \mathcal{P}_j^{(d)}, \quad (63)$$

are remapped back to the (Eulerian) reference levels given in (54) every `se_rspl` dynamics time-steps [Lin, 2004]. In the remapping process we enforce conservation of mass by mapping $m_k^{(\ell)} \Delta \mathcal{P}_k^{(d)}$ using the piecewise-parabolic method [PPM; Colella and Woodward, 1984] and applying a standard shape-preserving limiter to avoid unphysical (in particular negative) mixing ratios in the remapping process. The internal energy is also conserved during the remapping process by mapping $\sum_{\ell \in \mathcal{L}_{all}} c_p^{(\ell)} m_k^{(\ell)} T_k \Delta \mathcal{P}_k^{(d)}$. Note that temperature must be recovered from the internal energy using the remapped tracer values for $m_k^{(\ell)}$. A shape-preserving filter is also used for the remapping of internal energy so that an isothermal profile remains isothermal if the limiter is active on one of the water variables, $m_k^{(\ell)}$. Note that if the limiter is active at the same points for more than one of the water species then one can not guarantee preservation of an isothermal atmosphere when non-linear limiting filters are turned on [see, e.g., Section 2.5 in Lauritzen and Thuburn, 2012].

The moist mass-weighted velocity components, $\sum_{\ell \in \mathcal{L}_{all}} m_k^{(\ell)} \Delta \mathcal{P}_k^{(d)} u_k$ and $\sum_{\ell \in \mathcal{L}_{all}} m_k^{(\ell)} \Delta \mathcal{P}_k^{(d)} v_k$ respectively, are remapped separately. Mapping the moist mass weighted velocity components conservatively leads to an AAM conserving vertical remapping algorithm. For the velocity components we do not enforce shape-preservation in the vertical remapping process to reduce the amount of kinetic energy dissipation.

3.2 Horizontal discretization using the SE method

The CAM-SE uses a cubed-sphere geometry originally introduced by Sadourny [1972] to represent the planet earth. The spherical surface \mathcal{S} is a patched domain, which is partitioned into non-overlapping quadrilateral elements Ω_e such that $\mathcal{S} = \cup \Omega_e$ (see Fig. 2). On \mathcal{S} each 2D element $\Omega_e(x^1, x^2)$ defined in terms of central (gnomonic) projection angles $x^1, x^2 \in [-\pi/4, \pi/4]$, which serve as the independent variables in the computational domain. The mapping from cube to sphere results in a non-orthogonal curvilinear coordinate system on \mathcal{S} , with the metric tensor G_{ij} and analytic Jacobian $\sqrt{G} = |G_{ij}|^{1/2}$, $i, j \in \{1, 2\}$. A physical vector quantity such as the wind vector $\mathbf{v} = (u, v)$, defined on \mathcal{S} in orthogonal lat-long coordinates, can be uniquely expressed in tensor form using conventional notations as the covariant (u_1, u_2) and contravariant (u^1, u^2) vectors using the 2×2 transformation matrix \mathbf{D} associated with the gnomonic mapping such that $\mathbf{D}^T \mathbf{D} = G_{ij}$ (see Nair et al. [2005] for the details):

$$\begin{bmatrix} u \\ v \end{bmatrix} = \mathbf{D} \begin{bmatrix} u^1 \\ u^2 \end{bmatrix} = \mathbf{D}^{-T} \begin{bmatrix} u_1 \\ u_2 \end{bmatrix}. \quad (64)$$

The governing equations defined in familiar vector form can also be expressed in general tensor form. In order to describe the SE discretization process in simple terms, we

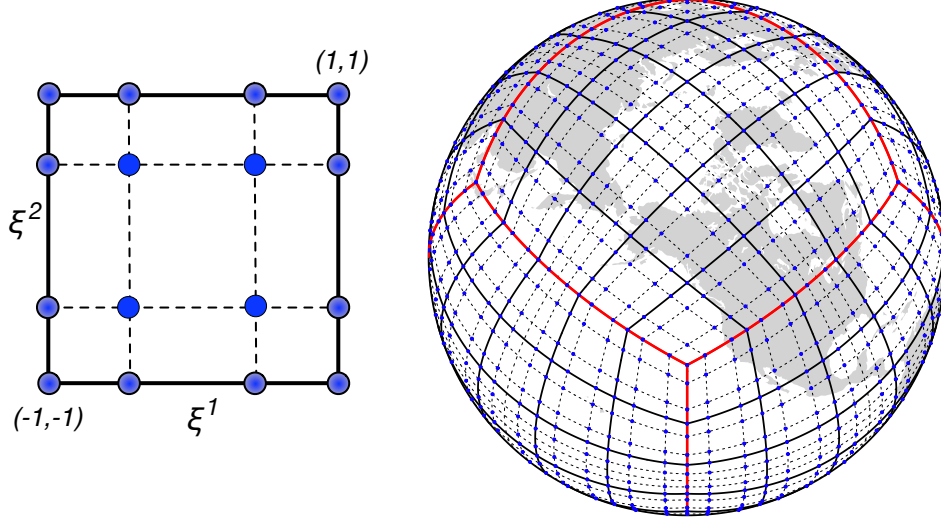


Figure 2. The left panel shows the Gauss-Lobatto-Legendre (GLL) grid with $N_p \times N_p$ quadrature points defined on a standard element $[-1, 1]^2$, where $N_p = 4$. The right panel shows the cubed-sphere (S) grid system tiled with $6N_e^2$ spectral elements Ω_e , where N_e is the number of elements in each coordinate direction on a panel. Each element Ω_e on S has the GLL grid structure.

consider the the following conservation law on S for an arbitrary scalar ϕ :

$$\frac{\partial \phi}{\partial t} + \nabla \cdot \mathbf{F}(\phi) = S(\phi) \quad \Rightarrow \quad \frac{\partial \phi}{\partial t} + \frac{1}{\sqrt{G}} \left[\frac{\partial \sqrt{G} F^1}{\partial x^1} + \frac{\partial \sqrt{G} F^2}{\partial x^2} \right] = S(\phi), \quad (65)$$

where the equation on the right side is a special case, the flux-form transport equation, with contravariant fluxes $(F^1, F^2) = (u^1 \phi, u^2 \phi)$, and $S(\phi)$ is an arbitrary source term.

3.3 SE spatial discretization in 2D

The SE solution process involves casting the partial differential equation in Galerkin form, i.e., by multiplying (65) with a test (weight) function ψ and integrating over the domain S ,

$$\int_S \psi \left[\frac{\partial \phi}{\partial t} + \nabla \cdot \mathbf{F}(\phi) - S(\phi) \right] dS = 0. \quad (66)$$

A computational form of (66) is obtained by applying Green's theorem, resulting in the weak Galerkin form, as follows:

$$\int_S \psi \frac{\partial \phi}{\partial t} dS = \int_S \nabla \psi \cdot \mathbf{F}(\phi) dS + \int_S \psi S(\phi) dS, \quad (67)$$

where the approximation to the solution ϕ and the test function belong to a polynomial space V^N . The SE method consists of partitioning the domain into non-overlapping elements and solving the global problem locally on each element, where the solution is approximated by using a set of basis (polynomial) functions of prescribed order N . A basic assumption used in SE (or continuous Galerkin) method is that the global basis corresponding to (67) is C^0 continuous. Therefore the problem (67) can be solved locally for each element Ω_e , if there is a mechanism by which the solution maintains C^0 continuity at the element boundaries as required by the SE discretization.

For efficient evaluation of the integral equation (67), the SE method employs the Gauss-Lobatto-Legendre (GLL) quadrature rule for integrals and collocation differentiation

for derivative operators. All the corresponding numerical operations are performed on a square $[-1, 1]^2$ known as the standard (or reference) element. In order to facilitate local mesh refinement, the spectral elements Ω_e on \mathcal{S} are defined as arbitrary spherical quadrilaterals in the CAM-SE grid system, which should be mapped onto the standard element. A direct way to address this problem is establishing a transformation $\mathcal{J}_e : \Omega_e \rightarrow [-1, 1]^2$, where \mathcal{J}_e may be considered as a composite mapping combining the gnomonic and the quadrilateral to standard-element mapping. Let the Jacobian associated with the composite mapping be $J_e = J_e(\sqrt{G})$. Then an arbitrary surface integral on Ω_e can be expressed in terms of local coordinates $\xi^1, \xi^2 \in [-1, 1]$ and the Jacobian J_e

$$\int_{\Omega_e} \psi(x^1, x^2) d\Omega_e = \int_{-1}^1 \int_{-1}^1 J_e(\xi^1, \xi^2) \psi(\xi^1, \xi^2) d\xi^1 d\xi^2 \approx \sum_{k=0}^N \sum_{l=0}^N w_k w_l J_e(\xi_k^1, \xi_l^2) \psi(\xi_k^1, \xi_l^2), \quad (68)$$

where w_k, w_l are the Gauss quadrature weights.

In the case of GLL quadrature rule, the nodal points $\xi_k, k = 0, 1, \dots, N$, are the roots of the polynomial $(1 - \xi^2)P'_N(\xi) = 0, \xi \in [-1, 1]$; and the corresponding GLL quadrature weights are given by

$$w_k = \frac{2}{N(N+1)[P'_N(\xi_k)]^2},$$

where $P_N(\xi)$ is the Legendre polynomial of degree N . For the SE discretization it is customary to use Lagrange polynomials $h_k(\xi)$, with roots at the GLL quadrature points ξ_k , as basis functions. This setup provides discrete orthogonality for the basis function $h_k(\xi)$, which is formally defined as:

$$h_k(\xi) = \frac{(\xi^2 - 1) P'_N(\xi)}{N(N+1) P_N(\xi_k) (\xi - \xi_k)}. \quad (69)$$

Note that there are $N+1 = N_p$ GLL quadrature points in 1D, and $N_p \times N_p$ GLL points are needed for 2D spectral elements Ω_e . Figure (2) shows the GLL grid with $N_p = 4$ on the left panel, and the right panel shows the cubed-sphere grid \mathcal{S} tiled with elements Ω_e , each with the GLL grid points.

A semi-discrete form of (67) on an element Ω_e can be obtained by approximating the solution as a tensor product of 1D Lagrange basis $\{h_k(\xi)\}_{k=0}^N$ such that

$$\phi|_{\Omega_e} \approx \phi^e(\xi^1, \xi^2, t) = \sum_{k=0}^N \sum_{l=0}^N \phi_{kl}^e(t) h_k(\xi^1) h_l(\xi^2), \quad (70)$$

where $\phi_{kl}^e(t) = \phi^e(\xi_k^1, \xi_l^2, t)$ are the nodal grid-point values of the solution, and defining the test function as $\psi(\xi^1, \xi^2) = h_k(\xi^1) h_l(\xi^2)$. By using (68) and the discrete orthogonality property of $h_k(\xi)$, we get a completely decoupled system of ODEs on Ω_e , for each grid-point (k, l)

$$M_{kl}^e \frac{d}{dt} \phi_{kl}^e(t) = A_{kl}^e + S_{kl}^e \quad (71)$$

$$M_{kl}^e = \int_{-1}^1 \int_{-1}^1 J_e h_k(\xi^1) h_l(\xi^2) d\xi^1 d\xi^2 = J_e(k, l) w_k w_l \quad (72)$$

$$A_{kl}^e = \sum_{i=0}^N J_e^{(1)}(i, l) F_{il}^1 D_{ik}^{(1)} w_i w_l + \sum_{i=0}^N J_e^{(2)}(k, i) F_{ki}^2 D_{li}^{(2)} w_k w_i \quad (73)$$

$$S_{kl}^e = J_e(k, l) w_k w_l S(U_{kl}) \quad (74)$$

where $J_e^{(i)} = J_e \partial \xi^i / \partial x^i$ is the metric term and $D_{lk}^{(i)}$ is the derivative matrix $h'_k(\xi_l^i)$, along the x^i -direction and $i \in \{1, 2\}$. The ODEs (71) can be written in a formal matrix-vector form for Ω_e following Karniadakis and Sherwin [2013]:

$$\mathbf{M}^e \frac{d}{dt} \Phi^e = \mathbf{A}^e + \mathbf{S}^e + \mathbf{B}^e, \quad (75)$$

where \mathbf{M}^e is the so-called mass matrix, which is a diagonal matrix with entries M_{kl}^e . \mathbf{B}^e indicates the boundary terms for the element Ω_e , which is a key component linking the local and global problem (67) and enforcing C^0 continuity for solutions across element boundaries.

The global matrices associated with (67) can be obtained by summing the contributions from elemental matrices and this procedure is known as the direct stiffness summation (DSS). However the global matrices are not explicitly constructed. In practice, the DSS operation replaces interface values of two contiguous elements sharing the same physical location by the weighted sum (average) so that the boundary nodes get unique values, which maintains the continuity of the global solution across the element edges. This strategy has been adopted in CAM-SE. Note that the DSS operation does not affect interior nodal values of any element, and preserves global conservation of the SE discretization for (65). The elemental discretization (75) combined with the DSS operation leads to the time-dependent system of ODE corresponding to (65),

$$\frac{d}{dt}\phi(t) = DSS(\phi). \quad (76)$$

The details of the discretization of the dissipation operators are provided in Appendix A: .

3.4 Mimetic discretization

A numerical method is mimetic (or compatible) if key integral properties of divergence, gradient and curl-operators are mimicked in discretized space. The CAM-SE discretization satisfies the divergence/gradient adjoint relation

$$\int \phi \nabla \cdot \vec{v} dS + \int \vec{v} \nabla \phi dS = 0 \quad (77)$$

in discretized space [Taylor and Fournier, 2010]. This property can be used to show the inherent conservation properties of CAM-SE in terms of mass and energy in the horizontal discretization. This is discussed in detail in Taylor [2011] and hence not repeated here.

Since the CAM-SE discretization is mimetic the adiabatic, frictionless discretization of the equations of motion conserves total moist energy to within time-truncation errors. The equivalent internal energy change due to hyperviscous damping of the velocity vector is added as frictional heating in the thermodynamic equation so the energy budget for the viscosity on the momentum equations is closed. The dissipation of energy due to hyperviscosity on temperature and pressure, however, is not energy-conserving. The vertical remapping does not conserve total energy either but it does conserve internal energy and AAM.

3.5 Temporal discretization

In a typical CAM setup where the model top is around 40km the maximum stable time-step for solving the continuity equation for water species ℓ , $\ell \in \mathcal{L}_{water}$ (referred to as tracer advection), is limited by the maximum advective winds. The maximum stable time-step for the remaining equations of motion (thermodynamic equation, momentum equations and the continuity equation for dry air) is limited by the fastest gravity waves. Since the maximum advective winds are typically slower than the fastest gravity waves, different time-stepping methods are used for tracer advection and the remaining equations of motion. While the overall time-step is the same, different Runge-Kutta (RK) methods are used to maintain stability.

The SE tracer advection algorithm uses a three-stage RK strong-stability-preserving (SSP) time-stepping method, ensuring the time step will preserve any shape-preserving

Table 1. The table shows physics time-step (column 3), sub-cycling parameters (columns 4-6) as a function of resolution (column 1-2). The horizontal resolution is specified in the format $ne30np4$ which refers to $N_e = 30$ and $N_p = 4$; similar for other resolutions. The approximate average grid-spacing, Δx_{ave} (in kilometers), is also listed (column 2). **Colin: please double-check CONUS variables**

Horizontal res.	Δx_{ave}	Δt_{phys}	se_nsplrit	se_rsplrit	se_hypervis_subcycle
ne16np4	~220km	1800s	1	3	1
ne30np4	~110km	1800s	2	3	2
ne60np4	~56km	1800s	4	3	2
ne120np4	~28km	600s	4	3	3
ne240np4	~14km	600s	6	3	3
ne0np4	~110km→~28km	?	5	3	1

properties preserved by the underlying spatial discretization [Spiteri and Ruuth, 2002]. The shape-preserving filter used is described in Guba *et al.* [2014b]. The shape-preserving SE tracer advection algorithm is formally second-order accurate.

The momentum, thermodynamic and dry air continuity equations are solved using a third-order accurate five-stage Runge-Kutta method (RK5) based on a modified version of Kinnmark and Gray [1984a,b] described in Guerra and Ullrich [2016]. The inviscid equations of motion are advanced one dynamics time-step using the RK5 method followed by an application of the hyperviscosity operators through solving the advection-diffusion equation of the updated prognostic variables. Note that it is necessary to subcycle the hyperviscosity-step for stability. The viscosity terms are computed as described in section 2.6.1 and discretized using the SE method presented in Appendix A: .

[Mark: How is tracer advection coupled to dry air continuity equation??]

The time-steps in CAM-SE are controlled with namelist variables `se_nsplrit`, `se_rsplrit`, `se_hypervis_subcycle` so that the time-steps for vertical remapping Δt_{remap} , tracer advection Δt_{trac} (RK3), RK5 time-stepping for non-tracers Δt_{dyn} , and hyperviscosity Δt_{hyper} are

$$\Delta t_{remap} = \frac{\Delta t_{phys}}{se_nsplrit}, \quad (78)$$

$$\Delta t_{dyn} = \frac{\Delta t_{phys}}{se_nsplrit * se_rsplrit}, \quad (79)$$

$$\Delta t_{hyper} = \frac{\Delta t_{phys}}{se_nsplrit * se_rsplrit * se_hypervis_subcycle}, \quad (80)$$

$$(81)$$

where Δt_{phys} is the time-step used for computing physics tendencies. For the 1° configuration ($N_e = 30$, $N_p = 4$ where N_e is the number of elements in each coordinate direction on a panel and N_p is the number of quadrature points along each coordinate direction in an element) the physics time-step is $\Delta t_{phys} = 30min$, and the subcycling parameters are `se_nsplrit=2`, `se_rsplrit=3` and `se_hypervis_subcycle=1`. See Table 1 for other resolutions.

3.6 Coupling to physics

CAM-SE uses a time-split approach in which dynamics advances the model state and the physics tendencies are based on the dynamics updated state. When CAM-SE starts a simulation, the physics are called first and then the dynamics. The question is then how

to add the physics tendencies in the dynamical core. CAM-SE supports several physics-dynamics coupling methods. Let F_X be the physics tendency for prognostic variable X in level k (for notational simplicity the vertical index is dropped). The different coupling methods are detailed below and are identified with the rather arbitrary name ‘ftype’. A fuller discussion with results about different coupling methods is the content of a separate paper. Some results are provided by C. Jablonowski in *Gross et al.* [2016, ; see section 6].

3.6.1 ‘ftype=0’ configuration (default)

In this case, compute the physics tendencies and add $\Delta t_{remap} F_X$ to the state of X , advance the dynamics Δt_{remap} seconds based on the updated state, add $\Delta t_{remap} F_X$ to the dynamics updated state of X and advance the dynamics core, and so on. In other words, the forcing is split into $\frac{\Delta t_{phys}}{\Delta t_{remap}}$ equal chunks and added throughout the dynamics. The ‘ftype=0’-option is used by default in CAM-SE and used for all simulations presented in this paper.

The CAM parameterization package returns mixing ratio tendencies for tracers. We convert the mixing ratio tendencies to mass tendencies $F_x \Delta \mathcal{P}_k^{(d)}$ since the prognostic variable for tracers in the dynamical core is $\Delta p_k m_k^{(d)}$.

Generally modelers do not allow the tracer tendencies to drive the mixing ratio of tracers negative, however, this may happen using ‘ftype=0’. If the tendency drives a mixing ratio negative the mixing ratio is set to zero. In this case, only a fraction of the entire tracer tendency is added leading, to an inconsistency in how much mass the physics package wants to remove and the amount of tracer mass actually removed in the physics-dynamics coupling code in the dynamical core.

3.6.2 ‘ftype=1’ configuration

In this configuration the entire physics forcing is added to the dynamics state $\Delta t_{phys} F_X$ which is equivalent to having the physics module update the model state. This coupling method is used in CAM-FV [Lin, 2004]. Note that contrary to the ‘ftype=0’ configuration, this configuration always provides a closed mass budget for tracers in terms of physics tendencies being fully applied in the dynamical core. The disadvantage of using ‘ftype=1’ in CAM-SE is that the use of a long physics time-step results in spurious gravity waves appear due to large physics tendencies.

3.6.3 ‘ftype=2’ configuration

This configuration constitutes a hybrid approach where mass variables (tracers) use the ‘ftype=1’ physics-dynamics coupling method and all other variables use the ‘ftype=0’ method.

4 Results

The evaluation of the SE dynamical core with CAM6 physics in an AMIP configuration simulation is the subject of a separate paper. Here the new dynamical core is evaluated in simpler configurations. First the new dynamical core version is compared with the old version using an idealized moist baroclinic wave thereby avoiding the complexity of a full physics parameterization suite that may make it harder to distinguish between cause and effect. It is confirmed that the new SE dynamical core version converges to within the uncertainty of high resolution reference solutions computed with the old SE dynamical core (that has been extensively validated) and the current ‘workhorse’ dynamical core for 1° climate simulation; CAM-FV [finite-volume Lin, 2004]. Second, the dynamical core is evaluated in an aqua-planet setup [Neale and Hoskins, 2000; Williamson et al., 2012; Medeiros et al., 2016] using CAM6 physics. In this context the conservation properties

AAM and total moist energy is discussed. Thirdly the computational efficiency of CAM-SE is evaluated.

4.1 Idealized moist baroclinic wave with Kessler microphysics

For the validation of the new dynamical core version in a simplified setup, we use a moist variant of the dry baroclinic wave of *Ullrich et al.* [2014] with Kessler microphysics [Kessler, 1969]. This test case configuration was part of the Dynamical Core Model Inter-comparison Project (DCMIP) 2016 test case suite [Ullrich et al., 2017]. The initialization of the atmospheric state for the moist baroclinic wave is based on analytic expressions for T_v , \vec{v} , p and $q^{(wv)}$ as a function of latitude and height; (φ, z) . The surface geopotential is constant, $\Phi_s = 0$, and the moist surface pressure is constant $p_s = 1000hPa$. The analytical expressions for temperature, velocity components, moist pressure and specific humidity, denoted $T_v(\varphi, z)$, $\vec{v}(\varphi, z)$, $p(\varphi, z)$, and $q^{(wv)}(\varphi, z)$, respectively, are given in Appendix C: . When using a moist vertical pressure coordinate, the full level pressures p_k of the initial condition are known from the hybrid coefficients, A_k and B_k , and one can iteratively solve for z_k given moist pressure: $p_k = p(\varphi, z_k)$ [see, Ullrich et al., 2014]. Once the full level heights, $z_k(\varphi)$, are known then the specific humidity, virtual temperature and velocity components can be computed by evaluating the analytical expressions at (φ, z_k) . In the DCMIP 2016 test case documentation the virtual temperature is converted to temperature using $T_v = T(1 + \epsilon q^{(wv)})$. For a dry-mass vertical coordinate model the initialization procedure is more complicated as described below.

4.1.1 Initialization of the moist baroclinic wave using a dry-mass vertical coordinate

The challenge is to extract dry pressure from the initial state defined in terms of moist pressure and preserve a balanced initial condition. For that we write the dry atmosphere hydrostatic relation (35) in terms of specific humidity, virtual temperature and moist pressure

$$\frac{\partial \mathcal{P}^{(d)}}{\partial z} = -\rho_d g, \quad (82)$$

$$= -\frac{\rho}{1 + m^{(wv)}} g, \quad (83)$$

$$= -\frac{p}{R^{(d)}T_v} \frac{1}{(1 + m^{(wv)})} g, \quad (84)$$

$$= -\frac{p}{R^{(d)}T_v} (1 - q^{(wv)}) g, \quad (85)$$

where we have substituted the ideal gas law and used $\left(\frac{1}{1+m^{(wv)}}\right) = (1 - q^{(wv)})$ for an atmosphere not containing any condensates. Hence the dry pressure as a function of latitude, φ , and height, z , is

$$\mathcal{P}^{(d)}(\varphi, z) = p_t - \int_z^{z_t} \frac{p(\varphi, z)}{R^{(d)}T_v(\varphi, z)} (1 - q^{(wv)}(\varphi, z)) g dz, \quad (86)$$

where z_t is the height of the model top computed by iteratively solving (C.4) with $p(\varphi, z) = p_t$ (assuming there is no moisture above the model top). To initialize the dry pressure levels a dry surface pressure is needed. It is computed by integrating (86) with lower integral bound $z = z_s = 0m$ and approximating the integral using 20 point Gaussian quadrature. The dry surface pressure is shown on Figure 3. The full dry pressure levels are given in terms of the hybrid coefficients $\mathcal{P}_k^{(d)}(\varphi, z_k) = A_k p_t + B_k \mathcal{P}_s^{(d)}(\varphi, z_k)$. The height, $z_k(\varphi)$, of the full dry pressure levels are computed by iteratively solving (86) with $\mathcal{P}^{(d)}(\varphi, z) = \mathcal{P}_k^{(d)}(\varphi, z_k)$. Once the heights are known the virtual temperature and velocity components can be computed by evaluating the analytical expressions at (φ, z_k) as is done for the moist vertical coordinate initialization. If we do the same for specific humidity then the moist surface pressure (which is a diagnostic when using dry-mass vertical coordinates) deviates more than 1 Pa in the tropics from the analytical value of 1000hPa (see

Figure 3). As a result a spurious zonal signal in surface pressure with the same amplitude as the initial gravity waves appears in the simulations. Hence specific humidity must be initialized more carefully in order to obtain a more balanced initial condition which is discussed in the next paragraph.

As for $\mathcal{P}^{(d)}$, the weight of water vapor $\mathcal{P}^{(wv)}$ per unit area can be written as

$$\mathcal{P}^{(wv)}(\varphi, z) = - \int_z^{z_t} \frac{p(\varphi, z)}{R^{(d)}T_v(\varphi, z)} q^{(wv)}(\varphi, z) g dz, \quad (87)$$

so that the moist pressure (in the absence of condensates) at half levels can be computed as the sum of dry air and water vapor pressures

$$p(\varphi, z_{k+1/2}) = \mathcal{P}^{(d)}(\varphi, z_{k+1/2}) + \mathcal{P}^{(wv)}(\varphi, z_{k+1/2}), \quad (88)$$

where $z_{k+1/2}$ is computed by the same iterative procedure as for full levels. An integrated value for water vapor in a layer can now be computed from

$$m^{(wv)}(\varphi, z_k) = \frac{p(\varphi, z_{k+1/2}) - p(\varphi, z_{k-1/2})}{\mathcal{P}^{(d)}(\varphi, z_{k+1/2}) - \mathcal{P}^{(d)}(\varphi, z_{k-1/2})} - 1. \quad (89)$$

Using this method for initializing the mixing ratio for water vapor the moist surface pressure is within 0.01 Pa of the analytical value of 10^5 Pa (see Figure 3). Once $m^{(wv)}(\varphi, z_k)$ is computed, the temperature $T(\varphi, z_k)$ can recovered from $T_v(\varphi, z_k)$ by using (17).

4.1.2 Simulation results

As part of the ‘CESM simpler models’ effort started by *Polvani et al.* [2017], the moist baroclinic wave setup has been implemented rigorously in the CESM in the sense that the configuration easily runs from CESM without code configurations using the ‘**FKESSLER** compset’. For instructions on how to run the moist baroclinic wave with Kessler micro-physics see *Lauritzen and Goldhaber* [2017]. Since the test case configuration has been implemented in the full CESM, the dynamical core interacts with the physics module as in full climate model simulations. Hence the global energy fixer is invoked [*Williamson et al.*, 2015] and for the dynamical cores using a moist pressure vertical coordinate there is an adjustment of specific humidity to conserve water after the moist physics updates [see Section 3.1.6 in *Neale et al.*, 2012]. The intent of this implementation is to evaluate the dynamical core in simplified setup but exactly as the dynamical core is configured for comprehensive climate simulations.

Figure 4 depicts day 1- pf the moist baroclinic wave test using the CAM-FV [Finite-Volume; *Lin*, 2004] dynamical core, CAM-HOMME dynamical core based on a moist vertical coordinate and the new dynamical core version (CAM-SE). The CAM-HOMME and CAM-SE (CESM2.0-SE) dynamical cores not only differ in terms of vertical coordinates but also in terms of hyperviscosity and the formula used for the heat-capacity in the thermodynamic equation. CAM-HOMME uses $c_p^{(d)}$ whereas CAM-SE (CESM2.0-SE) uses the comprehensive formula (24) that includes the heat capacity of water vapor.

To provide a more quantitative measure of the difference between the moist baroclinic wave simulations, the l_2 difference norm of p_s between two simulations is computed as the time varying global integral in spherical coordinates:

$$l_2(p_s(t)) = \left[\frac{1}{4\pi} \int_0^{2\pi} \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} (p_{s1}(\lambda, \varphi, t) - p_{s0}(\lambda, \varphi, t))^2 \cos(\varphi) d\varphi d\lambda \right]^{\frac{1}{2}}. \quad (90)$$

The l_2 difference norm between CAM-SE and CAM-HOMME is shown in Figure 5 for the 1° and $\frac{1}{4}^\circ$ resolution simulations. The l_2 norms for the 1° and $\frac{1}{4}^\circ$ simulations have

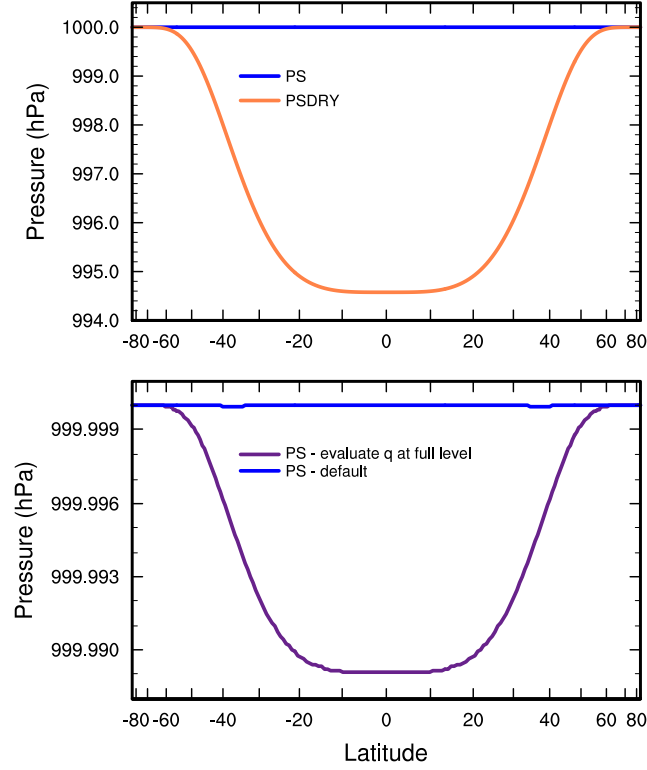


Figure 3. (Upper) Zonally averaged dry (orange) and moist (blue) surface pressure as a function of latitude for the numerically computed initial condition for the moist baroclinic wave at 1° horizontal resolution and 30 levels (CAM5 configuration). Due to increased water vapor towards the Equator the dry surface pressure decreases whereas the moist surface pressure is constant. (lower) Same as the upper plot but showing moist surface pressure where water vapor has been initialized by evaluating the analytic specific humidity formula at full levels (purple) and initializing the mixing ratio for water vapor in terms of moist and dry pressures at half levels which effectively integrates humidity over the layer (blue). Note that the upper and lower plots have different scales on the y-axis.

similar time varying magnitudes. As the baroclinic waves evolves, the l_2 norms grow to a maximum on the order of 1 hPa by day 15. To assess the significance of the l_2 norms, an l_2 norm that serves as an estimate of the uncertainty of a high resolution reference simulation is computed following *Jablonowski and Williamson* [2006]. The uncertainty in the reference is taken as the l_2 between a pair of $\frac{1}{4}^\circ$ resolution moist baroclinic wave simulations using different dynamical cores, CAM-SE and CAM-FV. At a $\frac{1}{4}^\circ$ resolution, the moist baroclinic wave solutions are converged to within a tolerable level of error (not shown), and therefore comparison between dynamical cores serves as an estimate of the uncertainty in the reference solutions arising from model imperfections.

The black curve in Figure 5 is the l_2 difference norm between CAM-SE and CAM-FV at approximately a $\frac{1}{4}^\circ$ resolution. l_2 values that fall below the uncertainty in the reference are considered insignificant. The l_2 between CAM-SE and CAM-HOMME generally lie near or below the uncertainty in the reference solution, indicating the differences between CAM-SE and CAM-HOMME are insignificant for the case of the moist baroclinic wave. This is consistent with the corresponding time varying, global minimum in p_s for all the simulations, which illustrates that the differences between CAM-SE and CAM-HOMME are smaller than the differences arising from increasing the horizontal resolution.

Day 10 moist baroclinic wave

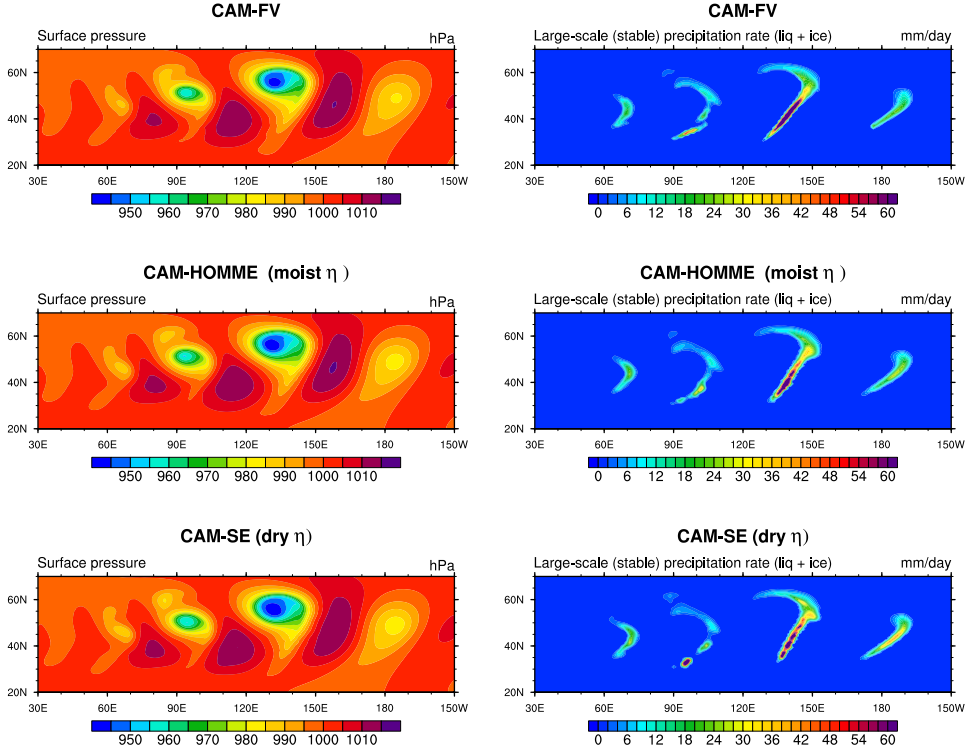


Figure 4. Left column shows moist surface pressure at day 10 for the moist baroclinic wave test case for (row 1) the finite-volume dynamical core, (row 2) CAM-HOMME version of the SE dynamical core based on a moist pressure vertical coordinate and (row 3) the dry-mass vertical coordinate version of SE presented in this paper. The right column is the same as the left but for large-scale precipitation rate.

4.2 CAM6 aqua-planet simulations

The 'CESM simpler models' effort also supports aqua-planet configurations [Medeiros *et al.*, 2016]. The aqua-planet configurations [Neale and Hoskins, 2000] refer to an ocean covered planet with no axial tilt – a planet in a perpetual equinox and devoid of continents. The aqua-planet compsets designate a fixed, zonally-symmetric SST distribution, modeled after the present day SST distribution on Earth ('QOBS' in Neale and Hoskins [2000]). The lack of a seasonal cycle allows one to compute robust statistics from a shorter simulation, and the absence of land removes any influence of discretized topography or interactions with the land model from the simulations. The aqua-planet configurations are therefore an indispensable tool for testing design choices in global atmospheric models.

To understand how the design choices adopted by CAM-SE influence the aqua-planet solutions compared to its predecessor, CAM-HOMME, we ran a pair of simulations using the CAM6 physics package [CAM6 reference] for 4.5 years. Figure 6 shows the total kinetic energy spectrum at the 200 hPa level in the two simulations. Compared to CAM-HOMME, the slope of the kinetic energy spectrum in CAM-SE is shallower for wavenumbers larger than 30, bringing the solutions closer to the empirically [Nastrom and Gage, 1985] and theoretically [Charney, 1971] determined slope of -3 at synoptic scales. The increased kinetic energy at smaller scales is primarily a result of the smaller divergence damping coefficient used in CAM-SE (not shown).

Figure 8b shows the zonally averaged total precipitation rate in CAM-HOMME (purple) and CAM-SE (red) averaged over the final 4 years of the simulations. The differences between the two simulations is provided as the purple curve on the bottom plot of Figure 8b. CAM-SE has increased precipitation at the equator, with a peak difference of 3 mm/day. An approximate 1 mm/day reduction in precipitation occurs on the flanks of the equator (Figure 8b). CAM-HOMME has twin Inter-Tropical Convergence Zones (ITCZ) straddling the equator, which appears absent from the CAM-SE simulations (Figure 8b). The total precipitation rate in the model is the sum of the precipitation due to the deep convective parameterization, and that due to CLUBB, which contains shallow convective precipitation and large-scale condensation. The deep convective precipitation rate indicates that the double-ITCZ does exist in CAM-SE (not shown), but is hidden from the total precipitation rate due to the increase in large-scale condensation at the equator.

An analysis of the vertically integrated, zonally averaged dry static energy budget indicates that the latent heating due to the change in precipitation rate are balanced by the anomalous dry static energy flux convergence due to the increase in the mean resolved vertical upward motion at the equator (the dynamic component from eqn. (3) in *Muller and O’Gorman* [2011]; not shown). This balance also holds for the reduction in precipitation rate on the flanks of the equator (not shown). It is likely that an increase in resolved vertical upward motion in CAM-SE drives the increase in large-scale condensation rate observed in CAM-SE, consistent with a prior analysis of CAM-HOMME [*Obrien et al.*, 2016]. In this scenario, greater downward motion observed on the flanks of the equator are simply compensating for the increased mass flux at the equator, which then acts to stabilize the column and reduce the precipitation rate locally.

The authors have identified three design aspects of CAM-SE that explain the large changes in precipitation observed in the Tropics. These three design aspects are: the use of a lower divergence damping coefficient, a thermodynamically consistent definition of c_p (see (24)) and the removal of the limiter in the vertical remapping of the horizontal winds. Figure 8b shows the total precipitation rate from three additional CAM-SE simulations, each simulation having one of the three modifications reverted back to the CAM-HOMME design. By using the larger divergence damping coefficients from CAM-HOMME in CAM-SE (‘CAM-SE-oldvisc’), the precipitation rate at the equator is reduced by about 1 mm/day compared with CAM-SE. The CAM-SE simulation in which the limiter is turned on (CAM-SE-ppmlimiter), as it is in CAM-HOMME, results in a dramatic increase in equatorial precipitation of 3–4 mm/day compared with CAM-SE (Figure 8b). Through reverting the definition of c_p back to the CAM-HOMME definition ($c_p = c_p^{(d)}$; ‘CAM-SE-cpcnst’), the near equatorial precipitation rates are dramatically reduced by 3–4 mm/day (Figure 8b). A forth simulation was performed containing all three of the aforementioned modifications (‘CAM-SE-all’). The simulated total precipitation rates in CAM-SE-all are indistinguishable from the CAM-HOMME simulation (Figure 8b).

The influence of the new definition of c_p and the PPM limiter on Tropical precipitation are not nearly as intuitive as the influence of the divergence damping coefficient. In our aqua-planet simulations, the ITCZ in part, consists of under-resolved, deep convective towers simulated by the resolved dynamics, consistent with a prior study of CAM-HOMME [*Herrington and Reed*, 2017]. An increase in divergence damping reduces the magnitude of vertical pressure velocities, likely explaining the reduction in precipitation in that simulation. The base of the convective towers often originate within the boundary layer, with convergent flow driving resolved upward mass fluxes through the cloud base. In the CAM-SE-ppmlimiter simulation, the magnitude of the horizontal velocities in the lowest model level of the equatorial convergence zone are larger (Figure 8b). The authors speculate that the limiter will select a larger magnitude horizontal wind in the lowest model level in a convergent flow regime, resulting in greater convergence, vertical motion and therefore precipitation.

To explain how the limiter works, consider a vertical profile of the horizontal wind, u_k , in the aqua-planet simulations across the lowest model level, $k=1$, and the second lowest model level, $k=2$ (Figure 8a). In the equatorial regions, there is a monotonic increase in the magnitude of the wind with height, across the lowest three levels. If level 1 is undergoing horizontal mass convergence, the model level will be raised (Figure 8a). To find the lowest model level wind on the new grid, u_1 , one integrates the PPM subgrid reconstruction from the surface to within a region that is occupied by level $k=2$. If the limiter is on, it forces the contribution of the wind to u_1 from the $k=2$ layer to be equal to u_2 . Since the magnitude of the wind increases with height, the contribution of the wind in u_1 from the region occupied by level $k=2$ will always have a smaller magnitude than u_2 , which is computed by integrating to the top of the $k=2$ layer, where the winds are largest (Figure 8a). In this scenario, the limiter will result in larger magnitude winds in the lowest model level, facilitating greater mass convergence during the following time-step.

The influence of a thermodynamically consistent definition of c_p is to increase precipitation near the equator by 3-4 mm/day (Figure 8b). Note that the $c_p = c_p^{(d)}$ change affects not only the energy conversion term in the thermodynamic equation but also the frictional heating term and the vertical remapping of temperature. Additional aqua-planet simulations reveal two important reasons for this sensitivity. A simulation in which $c_p = c_p^{(d)}$ only in the frictional heating term (equation (48)), results in a reduction in equatorial precipitation of 1-2 mm/day (not shown). This result is intuitive, since the term $(\vec{v} \cdot \delta \vec{v})$ is always negative, the heating rate is inversely proportional to c_p . In CAM-SE, c_p is always greater than $c_p^{(d)}$, and so there is more frictional heating in CAM-SE than there is in CAM-HOMME. In the equatorial regions, additional heating likely facilitates buoyancy, which increases resolved vertical motion, leading to an increase in precipitation. Setting $c_p = c_p^{(d)}$ just in the energy-conversion term in the thermodynamic equation also leads to a reduction in precipitation rates. We speculate that this is due to the reduced forcing term in the T -equation (third term on left-hand side of (39)) that in turn leads to larger densities (reduced temperature leads to increased density through the ideal gas law all else being equal) that then reduce the pressure gradient force (fourth term on left-hand side of (38)) and therefore less convergence in the Equatorial ‘belt’ of precipitation.

4.2.1 Conservation

To assess the AAM conservation properties of the CAM-SE dynamical core, the diagnostics used in *Lauritzen et al. [2014]* are applied to the Aqua-planet simulation described in Section 4.2. In the dynamical core the column integrated wind and mass AAM are written to history tapes and the total integrals of AAM are computed as a post-processing step. The torques are obtained by subtracting the global integrals of AAM divided by the time-increment between the outputs of the AAM. For details on the discretized AAM diagnostics see *Lauritzen et al. [2014]*. These diagnostic outputs are part of the CESM2 release and are controlled with namelist variables.

As discussed in Section 2.7 the total AAM torque from the dynamical core, in the absence of topography, should be small compared to the torque from the parameterizations. Figure 9 shows that that is indeed the case for CAM-SE. The spurious torques from the dynamical core are approximately a factor 100 smaller than the physical torques from the parameterizations. These results are very similar to the dynamical core torques found in the dry Held-Suarez setup used in *Lauritzen et al. [2014]*.

Similarly to the AAM diagnostics, the total column-integrated moist energy is output at various places in the dynamical core and physics parameterizations. The dynamical core conserves the moist total energy, equation (52), to about $0.1W/m^2$. The frictional heating term described in section 2.6.2 is approximately $0.4W/m^2$ and hence an important term for total moist energy conservation. As mentioned in section 2.7 the CAM physics energy fixer enforces a different energy than the comprehensive moist energy. This incon-

sistency should be removed in future CAM versions, however this is not a trivial modification to the CAM physics package. The discrepancy between the two definitions of energy is approximately $0.5W/m^2$ [similar to what *Taylor, 2011*, found when just including the correct heat capacity for water vapor in the total energy equation]. A detailed energy analysis will be the subject of a future publication.

4.3 Performance

[John Dennis et al.]

5 Conclusions

The NCAR CESM2.0 version of CAM-SE is presented. This version uses a dry-mass vertical coordinate and has a comprehensive treatment of condensates both in the thermodynamic equation and momentum equations. The conservation properties in terms of total energy and axial angular momentum for this system of equations derived. The discretization of the equations of motion is explained in detail and is intended to serve as a comprehensive documentation for CAM-SE. Idealized simulations demonstrate the accuracy and conservation property of the numerical model. In particular, we show that the reduction in viscosity parameters greatly improved the total kinetic energy spectrum of CAM-SE and that the comprehensive treatment of moist thermodynamics and condensate loading significantly changes precipitation rates in aqua-planet simulations. Last but not least the CAM-SE model has been sped-up significantly compared to its predecessor CAM-HOMME.

A: Dissipation

A.1 Discretization

In the CAM-SE model explicit fourth-order hyperviscosity ($\nu \nabla^4 \phi$) used as the main stabilization mechanism. For scalar fields such as T, p, q etc., a scalar viscosity is applied, while for the horizontal momentum equations vector viscosity ($\nu \nabla^4 \mathbf{v}$) is employed. The coefficient of viscosity ν may be constant or a spatially varying quantity depending on the application. High-order viscosity operator $\nu \nabla^{2m} \phi$, $m = 2, 3, \dots$, for an arbitrary variable ϕ can be constructed by successively applying the basic Laplacian-type viscosity operator $\nabla^2(\cdot)$. In order to describe the SE discretization we consider the basic Laplacian without the coefficient of viscosity ν :

$$L(\phi) = \nabla^2 \phi. \quad (\text{A.1})$$

Integrating (A.1) over an element Ω_e with boundary Γ_e , using Greens method results in

$$\int_{\Omega_e} L \psi \, d\Omega_e = \int_{\Gamma_e} \phi \nabla \psi \, d\Gamma_e - \int_{\Omega_e} \nabla \psi \cdot \nabla \phi \, d\Omega_e \quad (\text{A.2})$$

For the continuous Galerkin (SE) method the boundary integral vanishes ($\psi = 0$ at the element boundaries) and the *rhs* simplifies to a surface integral. The tensor gradients in the integrand can be expressed in terms of its contravariant components (\tilde{F}^1, \tilde{F}^2), using (64) such that

$$\nabla \psi \cdot \nabla \phi = \tilde{F}^1 \frac{\partial \psi}{\partial x^1} + \tilde{F}^2 \frac{\partial \psi}{\partial x^2}, \quad \begin{bmatrix} \tilde{F}^1 \\ \tilde{F}^2 \end{bmatrix} = \mathbf{D}^{-1} \mathbf{D}^{-T} \begin{bmatrix} \partial \phi / \partial x^1 \\ \partial \phi / \partial x^2 \end{bmatrix}. \quad (\text{A.3})$$

Thus the discretization of the Laplacian for SE method can be obtained by simplifying the integral

$$\int_{\Omega_e} L \psi \, d\Omega_e = - \int_{\Omega_e} \nabla \psi \cdot \nabla \phi \, d\Omega_e = - \int_{\Omega_e} \left[\tilde{F}^1 \frac{\partial \psi}{\partial x^1} + \tilde{F}^2 \frac{\partial \psi}{\partial x^2} \right] d\Omega_e. \quad (\text{A.4})$$

As in the case of (71), the weak formulation (A.4) can be evaluated on the standard element using the polynomial approximations (70) for ϕ, ψ on Ω_e and GLL quadrature rule.

Further simplification of (A.4) leads to

$$L_{kl}^e = -(M_{kl}^e)^{-1} \left[\sum_{i=0}^N J_e^{(1)}(i, l) \tilde{F}_{il}^1 D_{ik}^{(1)} w_i w_l + \sum_{i=0}^N J_e^{(2)}(k, i) \tilde{F}_{ki}^2 D_{li}^{(2)} w_k w_i \right], \quad (\text{A.5})$$

where L_{kl}^e is the value of Laplacian term for a gridpoint (k, l) on Ω_e .

Note that in practice, the contravariant gradient terms in (A.3) are first computed using collocation differentiation, then the weak divergence of the gradients is computed; this is followed by a DSS operation which yields the discrete Laplacian ($L(\phi^e) \Rightarrow \text{div}(\text{grad}(\phi^e))$).

In CAM-SE the vector viscosity is handled using the vector Laplacian $\nu \nabla^2 \mathbf{v}$, which is consistent with the curvilinear formulation of the momentum equation. By using the vector identity $\nabla^2 \mathbf{v} = \nabla(\nabla \cdot \mathbf{v}) - \nabla \times (\nabla \times \mathbf{v})$, the coefficient of viscosity ν may be split into the viscosity corresponding to the divergence part ν_d and that for the vorticity part ν_v such that

$$\nu \nabla^2 \mathbf{v} = \nu_d \nabla(\nabla \cdot \mathbf{v}) - \nu_v \nabla \times (\nabla \times \mathbf{v}). \quad (\text{A.6})$$

The SE discretization of (A.6) is handled separately for each component using a vector test function $\vec{\psi}$ and the weak formulation:

$$\nu_d \int_{\Omega_e} \vec{\psi} \cdot \nabla(\nabla \cdot \mathbf{v}) d\Omega_e = -\nu_d \int_{\Omega_e} (\nabla \cdot \vec{\psi})(\nabla \cdot \mathbf{v}) d\Omega_e \quad (\text{A.7})$$

$$-\nu_v \int_{\Omega_e} \vec{\psi} \cdot \nabla \times (\nabla \times \mathbf{v}) d\Omega_e = -\nu_v \int_{\Omega_e} (\nabla \times \vec{\psi}) \cdot (\nabla \times \mathbf{v}) d\Omega_e, \quad (\text{A.8})$$

where the line integrals along the boundaries associated with weak formulation vanishes for SE discretization. The *rhs* of the above equations are converted into equivalent tensor form for the divergence and curl terms, and discretized as in the case of (A.5).

A.2 Reference pressure for pressure-level thickness damping

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A.3 Hyperviscosity coefficients

The following hyperviscosity coefficients are used in CAM-SE:

$$\nu_T = \nu_{vor} = 0.2 \times \left(\frac{9 \times 10^6}{N_e(N_p - 1)} \right)^3 \frac{m^4}{s}, \quad (\text{A.9})$$

$$\nu_p = \nu_{div} = 1.0 \times \left(\frac{9 \times 10^6}{N_e(N_p - 1)} \right)^3 \frac{m^4}{s}, \quad (\text{A.10})$$

where $N_p = 4$ in CAM-SE, and $N_e = 30$ and $N_e = 120$ for the 1° and $1/4^\circ$ horizontal resolution configurations. Note that mass-wind consistency may be violated if $\nu_p \neq \nu_{div}$.

In the top three levels (sponge layer) the prognostic variables are damped with a Laplacian damping (second-order viscosity)

$$\nu_T = \nu_{vor} = \nu_{div} = \nu_p = \begin{cases} 4 \times 2.5 \frac{m^2}{s} & , \text{ for } k = 1, \\ 2 \times 2.5 \frac{m^2}{s} & , \text{ for } k = 2, \\ 1 \times 2.5 \frac{m^2}{s} & , \text{ for } k = 3, \end{cases} \quad (\text{A.11})$$

for resolutions up to (and including) $N_e = 120$.

B: Global conservation

Below it is shown that the continuous equations of motion, that include the effect of condensates, conserve AAM and total energy. For the derivations first note that adding the

continuity equations $\ell \in \mathcal{L}_{all}$ (40), using hydrostatic balance (41) and ideal gas law (15), yields a moist air continuity

$$\frac{\partial}{\partial t} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \right] + \nabla_{\eta^{(d)}} \cdot \left[\vec{v} \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \right] = 0, \quad (\text{B.1})$$

or, equivalently, in Lagrangian form

$$\frac{d}{dt} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \delta A \right] = 0, \quad (\text{B.2})$$

where δA is a horizontal area of a Lagrangian air parcel so that $\nabla \cdot \vec{v} = \frac{1}{\delta A} \frac{d}{dt} (\delta A)$, and $\frac{d}{dt} = \frac{\partial}{\partial t} + \vec{v} \cdot \nabla$ is the material/total derivative.

B.1 Axial angular momentum (AAM)

The conservation law for angular momentum is derived in spherical coordinates for which the zonal momentum equation takes the form

$$\frac{du}{dt} = \frac{uv \tan \varphi}{r} + 2\Omega v \sin \varphi - \frac{1}{\rho r \cos \varphi} \frac{\partial p}{\partial \lambda}, \quad (\text{B.3})$$

where u and v are the zonal and meridional velocity components, respectively, φ is latitude and λ longitude, Ω rotation rate of Earth ($f = 2\Omega \sin \varphi$), and r is the mean radius of Earth.

The conservation law for AAM, $\mathcal{M} = (u + \Omega r \cos \varphi) r \cos \varphi$, can be conveniently derived using the Lagrangian form. Consider the material derivative of \mathcal{M} multiplied by the volume of a Lagrangian air parcel, $\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \delta A$

$$\frac{d}{dt} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \delta A (u + \Omega r \cos \varphi) r \cos \varphi \right]. \quad (\text{B.4})$$

Using the chain rule, continuity equation on the form (B.2), the equality $v = r \frac{d\varphi}{dt}$ and substituting (B.3), one can obtain an evolution equation for AAM

$$\frac{d}{dt} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \delta A \mathcal{M} \right] = - \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \frac{\partial p}{\partial \lambda}, \quad (\text{B.5})$$

or, equivalently, in Eulerian form

$$\frac{\partial}{\partial t} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \mathcal{M} \right] + \nabla_{\eta^{(d)}} \cdot \left[\vec{v} \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \mathcal{M} \right] = - \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \frac{\partial p}{\partial \lambda}. \quad (\text{B.6})$$

Repeatedly using the chain rule one can show that the right-hand side of (B.5) and (B.6) can be written as

$$- \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \frac{\partial p}{\partial \lambda} = - \frac{\partial}{\partial \lambda} \left[\left(\frac{\partial z}{\partial \eta^{(d)}} \right) p \right] + p \frac{\partial}{\partial \lambda} \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \quad (\text{B.7})$$

$$= - \frac{\partial}{\partial \eta^{(d)}} \left(\frac{\partial z}{\partial \lambda} p \right) + \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \frac{\partial p}{\partial \lambda} + p \frac{\partial}{\partial \lambda} \left(\frac{\partial z}{\partial \eta^{(d)}} \right), \quad (\text{B.8})$$

$$= - \frac{\partial}{\partial \eta^{(d)}} \left(\frac{\partial z}{\partial \lambda} p \right) + \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \frac{\partial p}{\partial \lambda} - \frac{\partial}{\partial \lambda} \left[p \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \right] - \frac{\partial p}{\partial \lambda} \left(\frac{\partial z}{\partial \eta^{(d)}} \right). \quad (\text{B.9})$$

$$= - \frac{\partial}{\partial \eta^{(d)}} \left(\frac{\partial z}{\partial \lambda} p \right) - \frac{\partial}{\partial \lambda} \left[p \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \right]. \quad (\text{B.10})$$

Substituting (B.10) on the right-hand side of (B.6) and integrating (B.6) in the horizontal and vertical we results in the final AAM equation (50) in the main text.

B.2 Total energy

The derivation of the total energy equation closely follows *Kasahara* [1974] but for a dry mass vertical coordinate and inclusion of condensates in the equations of motion. The equation for the horizontal kinetic energy per unit mass, $K = \frac{1}{2} \vec{v} \cdot \vec{v}$, is derived by multiplying the momentum equations (38) with $\rho \vec{v} \left(\frac{\partial z}{\partial \eta^{(d)}} \right)$ and the continuity equation (B.1) with K , adding the two resulting equations and simplifying using the chain rule yields

$$\frac{\partial}{\partial t} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) K \right] + \nabla_{\eta^{(d)}} \left[\rho \vec{v} \left(\frac{\partial z}{\partial \eta^{(d)}} \right) K \right] = - \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \vec{v} \nabla_{\eta^{(d)}} p - g \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \vec{v} \cdot \nabla_{\eta^{(d)}} z. \quad (\text{B.11})$$

For the derivation of a flux-form version of the total energy equation (that includes a geopotential term) we substitute

$$g \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \vec{v} \cdot \nabla_{\eta^{(d)}} z = \nabla_{\eta^{(d)}} \cdot \left[g z \vec{v} \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \right] - g z \nabla_{\eta^{(d)}} \cdot \left[\vec{v} \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \right], \quad (\text{B.12})$$

$$= \nabla_{\eta^{(d)}} \cdot \left[g z \vec{v} \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \right] - z \nabla_{\eta^{(d)}} \cdot \left[\vec{v} \left(\frac{\partial p}{\partial \eta^{(d)}} \right) \right], \quad (\text{B.13})$$

(where the hydrostatic balance equation (41) has been used on the right-hand side of (B.13)) on the right-hand side of (B.11) and rearrange terms

$$\frac{\partial}{\partial t} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) K \right] + \nabla_{\eta^{(d)}} \cdot \left[\rho \vec{v} \left(\frac{\partial z}{\partial \eta^{(d)}} \right) (K + g z) \right] = - \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \vec{v} \nabla_{\eta^{(d)}} p - z \nabla_{\eta^{(d)}} \cdot \left[\vec{v} \left(\frac{\partial p}{\partial \eta^{(d)}} \right) \right]. \quad (\text{B.14})$$

Now, by multiplying the thermodynamic equation (39) with $\rho \frac{\partial z}{\partial \eta^{(d)}} c_p$ and the continuity equation (B.1) with $c_p T$, adding the resulting equations and simplifying using the chain rule yields

$$\frac{\partial}{\partial t} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) c_p T \right] + \nabla_{\eta^{(d)}} \left[\rho \vec{v} \left(\frac{\partial z}{\partial \eta^{(d)}} \right) c_p T \right] = \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \omega. \quad (\text{B.15})$$

By using that $\omega \equiv \frac{\partial p}{\partial t} + \vec{v} \cdot \nabla_{\eta^{(d)}} p$, using the chain rule and the continuity equation (B.1), one can show that

$$\left(\frac{\partial z}{\partial \eta^{(d)}} \right) \omega = \frac{\partial}{\partial \eta^{(d)}} \left(z \frac{\partial p}{\partial t} \right) + \left(\frac{\partial z}{\partial \eta^{(d)}} \right) \vec{v} \nabla_{\eta^{(d)}} p + z \nabla_{\eta^{(d)}} \cdot \left[\vec{v} \frac{\partial p}{\partial \eta^{(d)}} \right]. \quad (\text{B.16})$$

Substituting (B.16) on the right-hand side of (B.15) and adding the resulting equation to (B.14), the two terms on the right-hand side of (B.14) cancel the two last terms on the right-hand side of (B.16) and we get the total energy equation

$$\frac{\partial}{\partial t} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) (K + c_p T) \right] + \nabla_{\eta^{(d)}} \left[\rho \vec{v} \left(\frac{\partial z}{\partial \eta^{(d)}} \right) (K + g z + c_p T) \right] = \frac{\partial}{\partial \eta^{(d)}} \left(z \frac{\partial p}{\partial t} \right). \quad (\text{B.17})$$

This equation is in the same form as (5.7) in *Kasahara* [1974] but ρ , c_p and p include the effect of condensates. Noting that

$$c_p T \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) = c_v T \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) + g z \rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right), \quad (\text{B.18})$$

where we have used the ideal gas law on the form (26), chain rule and hydrostatic equation on the form (41), the total energy equation can be written on the form

$$\frac{\partial}{\partial t} \left[\rho \left(\frac{\partial z}{\partial \eta^{(d)}} \right) (K + c_v T + g z) \right] + \nabla_{\eta^{(d)}} \left[\rho \vec{v} \left(\frac{\partial z}{\partial \eta^{(d)}} \right) (K + g z + c_p T) \right] = - \frac{\partial}{\partial \eta^{(d)}} \left(p \frac{\partial z}{\partial t} \right). \quad (\text{B.19})$$

As an aside it is noted that in a z -based vertical coordinate (for a moment assume $\eta^{(d)} \equiv z$), if integrating (B.19) in the vertical and using that z is constant at the model top (z_{top}) and surface (z_s) we get

$$\frac{\partial}{\partial t} \int_{z=z_s}^{z=z_{top}} (K + c_v T + g z) \rho dz + \nabla_z \cdot \int_{z=z_s}^{z=z_{top}} \vec{v} (K + g z + c_p T) \rho dz = 0. \quad (\text{B.20})$$

Note the clear separation of kinetic (K), potential (gz) and internal ($c_v T$) energy. Integrating (B.20) in the horizontal over the entire sphere the flux term drops out, and it is clear that the total energy is conserved for the frictionless and adiabatic system of equations.

For the global integral of the total energy equation in a dry-mass vertical coordinate, we substitute the hydrostatic relation on the form (41) into (B.17), integrate in the vertical and use that the pressure at the model top is constant,

$$\begin{aligned} \frac{1}{g} \frac{\partial}{\partial t} \int_{\eta=0}^{\eta=1} \left(\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} \right) \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) (K + c_p T) d\eta^{(d)} \\ + \frac{1}{g} \nabla_{\eta^{(d)}} \cdot \int_{\eta=0}^{\eta=1} \vec{v} \left(\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} \right) \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) (K + gz + c_p T) d\eta^{(d)} = -z_s \frac{\partial p_s}{\partial t}, \end{aligned} \quad (\text{B.21})$$

which can also be written as

$$\begin{aligned} \frac{1}{g} \frac{\partial}{\partial t} \int_{\eta=0}^{\eta=1} \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) \sum_{\ell \in \mathcal{L}_{all}} [m^{(\ell)} (K + c_p^{(\ell)} T + \Phi_s)] d\eta^{(d)} \\ + \frac{1}{g} \nabla_{\eta^{(d)}} \cdot \int_{\eta=0}^{\eta=1} \vec{v} \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) \sum_{\ell \in \mathcal{L}_{all}} [m^{(\ell)} (K + c_p^{(\ell)} T + gz)] d\eta^{(d)} = 0. \end{aligned} \quad (\text{B.22})$$

by expanding c_p using (24), re-arranging terms and using that $\Phi_s = gz_s$ is time-independent and that

$$p_s = \int_{\eta=0}^{\eta=1} \sum_{\ell \in \mathcal{L}_{all}} \left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) m^{(\ell)} d\eta^{(\eta)}. \quad (\text{B.23})$$

Note that the energy terms (inside square brackets) in (B.22) separate into all the components of moist air

$$\left(\frac{\partial \mathcal{P}^{(d)}}{\partial \eta^{(d)}} \right) \sum_{\ell \in \mathcal{L}_{all}} [m^{(\ell)} (K + c_p^{(\ell)} T + \Phi_s)]. \quad (\text{B.24})$$

Similarly for the flux term in the second square brackets on the left-hand side of (B.22). Integrating (B.22) over the entire sphere results in (52) given in the main text.

C: Analytical initial condition functions

In this section the analytical expressions for the moist baroclinic wave are given. The moist surface pressure is constant $p_s = 1000 \text{ hPa}$, meridional wind component is zero, $v(\varphi, z) = 0 \text{ m/s}$, and the surface geopotential is zero, $\Phi_s(\varphi, z) = 0 \text{ m}^2/\text{s}^2$. The reference virtual temperature is given by

$$T_v(\varphi, z) = \left\{ \mathcal{F}_1(z) - \mathcal{F}_2(z) \left[(\cos \varphi)^{\mathcal{K}} - \frac{\mathcal{K}}{\mathcal{K} + 2} (\cos \varphi)^{\mathcal{K}+2} \right] \right\}^{-1}, \quad (\text{C.1})$$

where

$$\mathcal{F}_1(z) = \frac{1}{T_0} \exp \left(\frac{\Gamma z}{T_0} \right) + \left(\frac{T_0 - T_P}{T_0 T_P} \right) \left[1 - 2 \left(\frac{zg}{bR^{(d)}T_0} \right)^2 \right] \exp \left[- \left(\frac{zg}{bR^{(d)}T_0} \right)^2 \right] \quad (\text{C.2})$$

$$\mathcal{F}_2(z) = \frac{(\mathcal{K} + 2)}{2} \left(\frac{T_E - T_P}{T_E T_P} \right) \left[1 - 2 \left(\frac{zg}{bR^{(d)}T_0} \right)^2 \right] \exp \left[- \left(\frac{zg}{bR^{(d)}T_0} \right)^2 \right], \quad (\text{C.3})$$

with $T_0 = \frac{1}{2}(T_E + T_P)$. Parameter $T_E = 310 \text{ K}$ is the temperature at the Equatorial surface, $T_P = 240 \text{ K}$ is the polar surface temperature, $\mathcal{K} = 3$ is the jet width parameter, $b = 2$ is the jet half-width parameter, and $\Gamma = 0.005 \text{ K/m}$ is the lapse rate.

To maintain hydrostatic balance, the pressure is given by:

$$p(\varphi, z) = p_0 \exp \left[- \frac{g}{R^{(d)}} (\mathcal{F}_3(z) - \mathcal{F}_4(z) \mathcal{I}_T(\varphi)) \right] \quad (\text{C.4})$$

where

$$\mathcal{F}_3(z) = \frac{1}{\Gamma} \left[\exp\left(\frac{\Gamma z}{T_0}\right) - 1 \right] + z \left(\frac{T_0 - T_P}{T_0 T_P} \right) \exp \left[- \left(\frac{zg}{bR^{(d)}T_0} \right)^2 \right] \quad (\text{C.5})$$

$$\mathcal{F}_4(z) = \frac{(\mathcal{K} + 2)}{2} \left(\frac{T_E - T_P}{T_E T_P} \right) z \exp \left[- \left(\frac{zg}{bR^{(d)}T_0} \right)^2 \right]. \quad (\text{C.6})$$

To define the zonal velocity component define the great circle distance between (λ, φ) and (λ_p, φ_p) :

$$R(\lambda, \varphi; \lambda_p, \varphi_p) = r \arccos \left(\sin \varphi \sin \varphi_p + \cos \varphi \cos \varphi_p \cos(\lambda - \lambda_p) \right). \quad (\text{C.7})$$

The zonal velocity component is

$$u(\varphi, z) = -\Omega r \cos(\varphi) + \sqrt{(\Omega r \cos(\varphi))^2 + r \cos(\varphi) U(\varphi, z) + u'(\lambda, \varphi, z)}, \quad (\text{C.8})$$

where the zonally symmetric part of the velocity field is given by

$$U(\varphi, z) = \frac{g\mathcal{K}}{r} \mathcal{F}_4(z) \left[(\cos \varphi)^{\mathcal{K}-1} - (\cos \varphi)^{\mathcal{K}+1} \right] T_v(\varphi, z), \quad (\text{C.9})$$

$r = 6371.22 \text{ m}$ is the mean radius of Earth and angular velocity is $\Omega = \frac{2\pi}{86164} \frac{1}{s}$ (denominator is length of day is seconds), and $u'(\lambda, \varphi, z)$ is the exponential bell-shaped perturbation to the zonally balanced velocity field

$$u'(\lambda, \varphi, z) = \begin{cases} U_p Z(z) \exp \left[- \left(\frac{R(\lambda, \varphi; \lambda_p, \varphi_p)}{r_p} \right)^2 \right], & \text{if } R(\lambda, \varphi; \lambda_p, \varphi_p) < r_p, \\ 0, & \text{otherwise,} \end{cases} \quad (\text{C.10})$$

where perturbation velocity is $U_p = 1 \text{ m/s}$, longitude/latitude of the zonal wind perturbation centerpoint is $(\lambda_p, \varphi_p) = (\pi/9, 2\pi/9) = (20^\circ \text{E}, 40^\circ \text{N})$, and

$$Z(z) = \begin{cases} 1 - 3 \left(\frac{z}{z_p} \right)^2 + 2 \left(\frac{z}{z_p} \right)^3, & \text{if } z \leq z_p, \\ 0, & \text{otherwise,} \end{cases} \quad (\text{C.11})$$

where $z_p = 15000 \text{ m}$ is the maximum height of the zonal wind perturbation. The specific humidity (moist mixing ratio for water vapor) is specified in terms of moist pressure (as the vertical variable)

$$q^{(wv)}(\lambda, \varphi, p) = \begin{cases} q_0 \exp \left[- \left(\frac{\varphi}{\varphi_w} \right)^4 \right] \exp \left[- \left(\frac{p-p_0}{p_w} \right)^2 \right], & \text{if } p > p_0/10, \\ q_t, & \text{otherwise.} \end{cases} \quad (\text{C.12})$$

where $p_w = 340 \text{ hPa}$ is a pressure width parameter, $q_0 = 0.018 \text{ kg/kg}$ is maximum specific humidity, $q_t = 1.0 \times 10^{-12} \text{ kg/kg}$ is specific humidity above artificial tropopause, $\varphi_w = 2\pi/9$ is the specific humidity latitudinal width parameter. In addition to water vapor the Kessler microphysics water species are cloud liquid, $m^{(cl)}$, and rain water, $m^{(rw)}$. Both are initialized to zero kg/kg.

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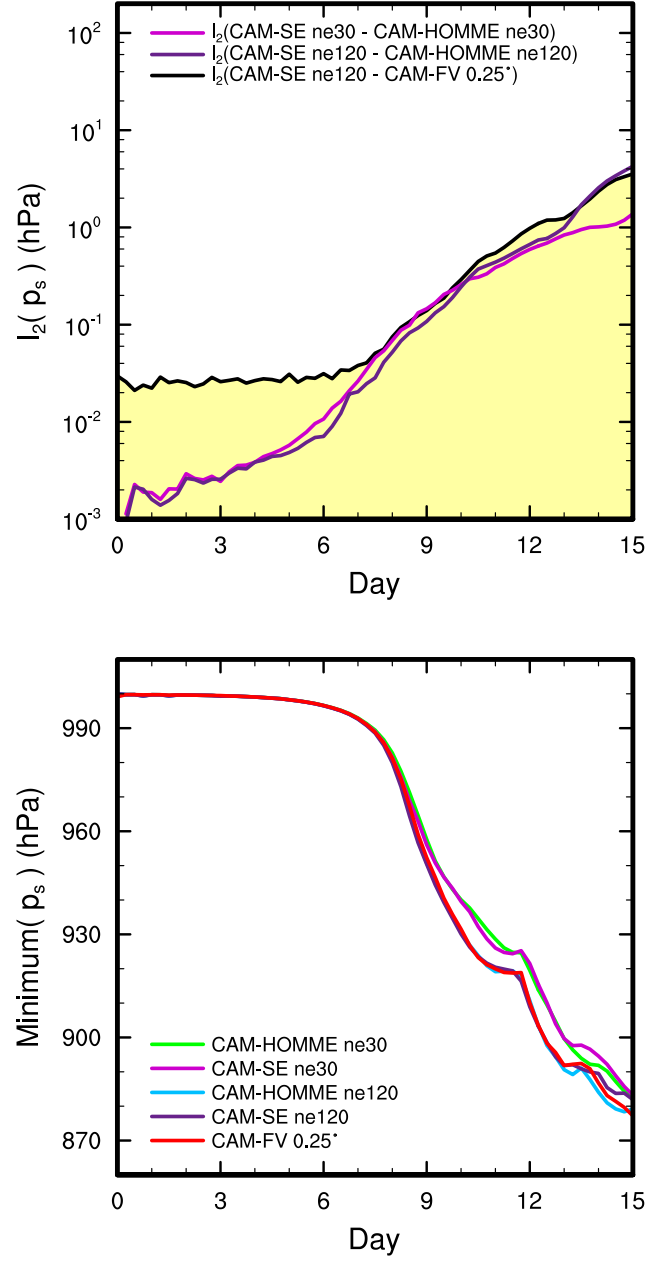


Figure 5. (Upper) l_2 difference norms of p_s in the moist baroclinic wave simulations. l_2 values lying within the yellow region fall below the estimate of the uncertainty in the reference solution (black curve). (lower) Global minimum p_s in the moist baroclinic wave simulations.

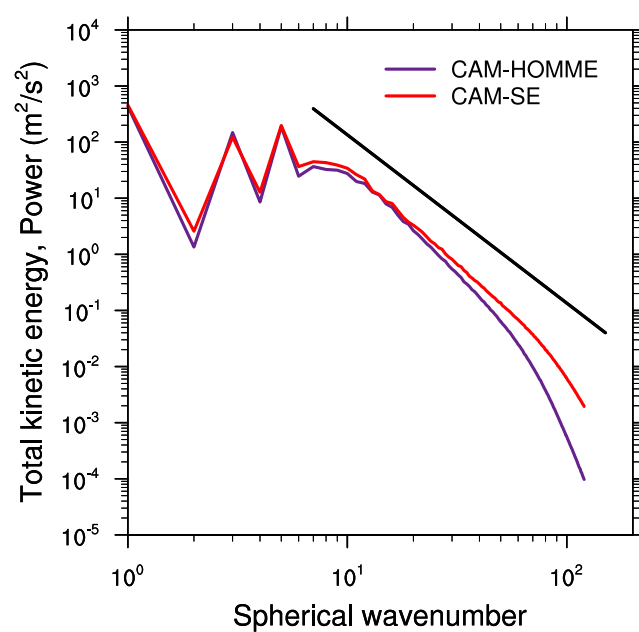


Figure 6. Total kinetic energy spectrum of the horizontal winds at the 200 hPa level in CAM-HOMME and CAM-SE, computed as the mean spectra 30 days of 6-hourly instantaneous spectra. Black line is the k^{-3} reference scaling.

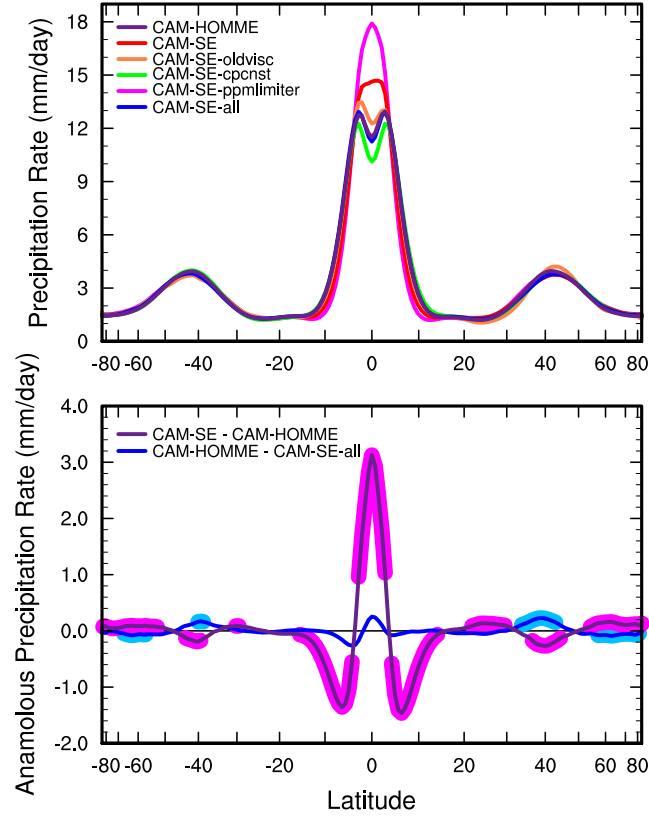


Figure 7. (Upper) The zonally average total precipitation rates in the aqua-planet simulations, averaged over the final 4 years of a 4.5 years simulation. Labels are defined in the text (lower) The change in the total precipitation rate between two simulations denoted by the label. The shading indicates where the differences are significant at the 95% confidence level.

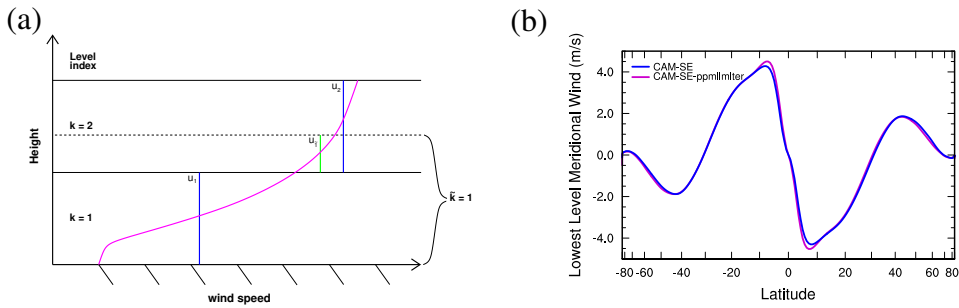


Figure 8. (a) A schematic illustration of the influence of the limiter on the vertical remapping of the winds of the lowest model level, in a convergent flow regime. Solid black lines show the layer interface, and the dashed line shows the top of the lowest model level in the following time-step. The model winds are denoted as blue-bars, while the PPM reconstruction is shown as the pink curve. The green bar is the wind computed by integrating the reconstruction from the top of level 1 to the top of level 1 in the following time-step. Notations are provided in the text. (b) Time-mean, zonally averaged lowest model level meridional wind in the CAM-SE and CAM-SE-ppmlimiter simulations, computed from the entirety of a 4.5 year simulation.

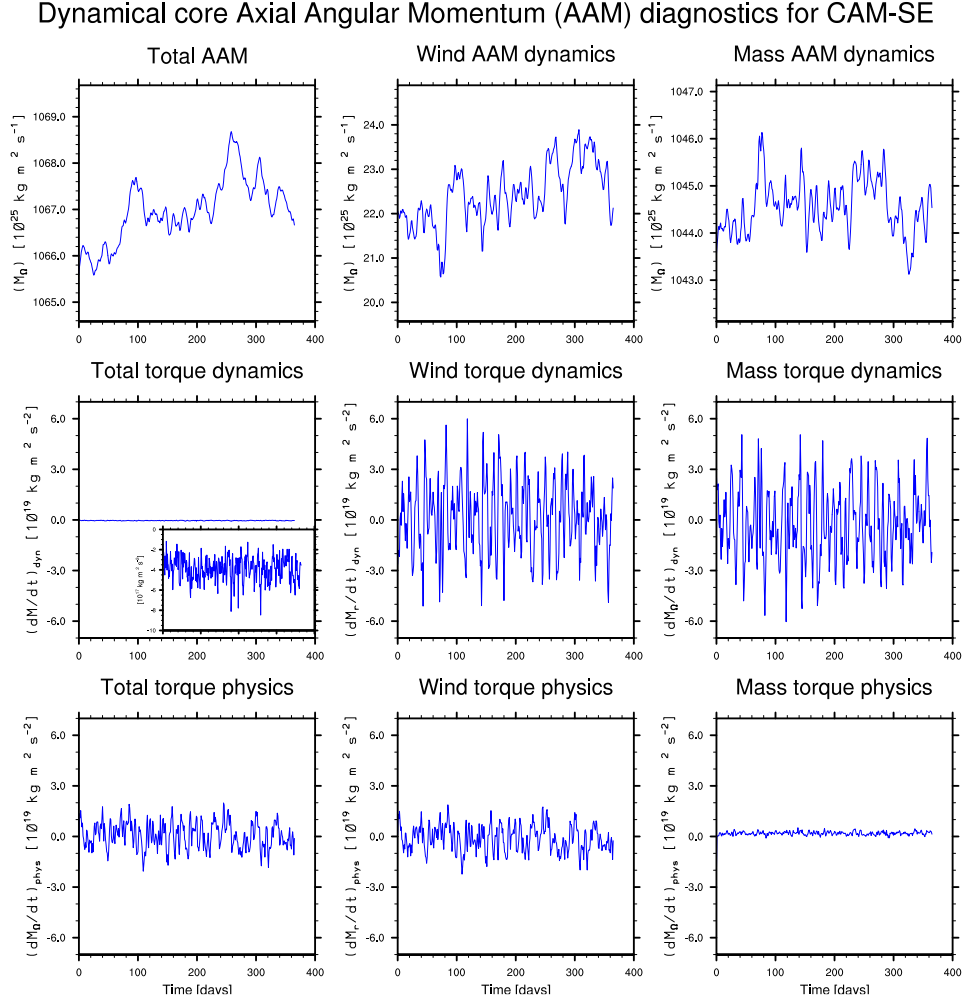


Figure 9. The first row shows total (column 1), wind (column 2) and mass (column 3) AAM as a function of time (**ADAM: is this daily instantaneous**) for the first year of the CAM6 aqua-planet simulation. The remaining plots show total, wind and mass AAM torques as a function of time (column 1,2,3, respectively) for the dynamical core (row 2) and parameterizations (row 3). Note that it is necessary to enlarge the y-axis of the dynamical core torque by a factor of approximately 100 to visualize the dynamical core torque. Hence the spurious torque from the dynamical core is small compared to the physical torques from the parameterizations.