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# The performance of the CoMorph-A convection package in global simulations with the Met Office Unified Model

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1 2 3 4 ORIGINAL ARTICLE 1 5 **Journal Section** 6 7 8 The performance of the CoMorph-A convection 9 10 package in global simulations with the Met Office 11 12 Unified Model 13 14 15 16 A. P. Lock<sup>1</sup> | M. Whitall<sup>1</sup> | A. J. Stirling<sup>1</sup> | 17 K. D. Williams<sup>1</sup> | S. L. Lavender<sup>2,1</sup> C. Morcrette<sup>1</sup> 18 19 6 K. Matsubayashi<sup>3</sup> | P. R. Field<sup>1</sup> G. Martin<sup>1</sup> Т 20 21 M. Willett<sup>1</sup> | J. Heming<sup>1</sup> 22 23 24 <sup>1</sup>Met Office, FitzRoy Way, Exeter, EX1 3PB. 25 26 <sup>2</sup>Centre for Applied Climate Sciences, The impact on global simulations of a new package of physical 27 University of Southern Queensland, parametrizations in the Met Office Unified Model is documented. Toowoomba, Australia. 28 The main component of the package is an entirely new convection <sup>3</sup>Japan Meteorological Agency, Minato City, 29 scheme, CoMorph. This has a mass-flux structure that allows initi-Tokyo, Japan 30 ation of buoyant ascent from any level and the ability for plumes of differing originating levels to coexist in a grid box. it has a different 31 Correspondence form of closure, where the mass-flux of initiation is dependent on 32 Met Office, FitzRoy Way, Exeter, EX1 3PB. local instability, and an implicit numerical solution for detrainment 33 Email: adrian.lock@metoffice.gov.uk that yields smooth timestep behaviour. The scheme is more con-34 sistently coupled to the cloud, microphysics and boundary layer **Funding information** 35 parametrizations and, as a result, significant changes to these have also been made. The package, called CoMorph-A, has been tested 36 in a variety of single-column and idealised regimes. Here we test it 37 in global configurations and evaluate it against observations using 38 a range of standard metrics. Overall it is found to perform well 39 against the control. Biases in the climatologies of the radiative fluxes are significantly reduced across the tropics and sub-tropics. 40 tropical and extratropical cyclone statistics are improved and the 41 MJO and other propagating tropical waves are strengthened. It 42 also improves overall scores in NWP trials, without revisions to the 43 data assimilation. There is still work to do to improve the diurnal 44 cycle of precipitation over land, where the peak remains too close to the middle of the day. 45 46 **KEYWORDS** 47 Convection, parametrization, global, evaluation, Unified Model 48 49 1 50 51 52 53

#### INTRODUCTION

The representation of convection in general circulation models (GCMs) has long been recognised as having a central control on model performance, not only influencing the distribution of precipitation, but also global circulation patterns, and the resulting cloud and radiative fluxes that respond to them (e.g. Xie et al. (2012), Holloway et al. (2014) and Sherwood et al. (2014)). Common to many GCMs, the Unified Model has previously used a mass-flux scheme based on simplified Arakawa-Schubert principles (Arakawa and Schubert (1974)), which was devised by Gregory and Rowntree (1990), and used largely predefined profiles for entrainment, a simple link to detrainment, along with a CAPE-based closure. While this scheme and its subsequent variants (e.g. Derbyshire et al. (2011), Walters et al. (2019), Willett and Whitall (2017)) has been able to produce good mean atmospheric states (e.g. Walters et al. (2019)), it has had difficulty generating appropriate coupling between convection and the model dynamics. This is seen in a range of emergent atmospheric phenomena that rely on the close feedbacks between convection and the larger-scale dynam-ics. Examples, in the UM and many other GCMs, include the insufficient strength and progression of the MJO, (e.g. Klingaman and Woolhough (2013) and Ahn et al. (2020)); the low amplitude of convectively-coupled Kelvin and other equatorial waves (Vitart et al. (2007), Janiga et al. (2018) and Dias et al. (2023)); the limited strength and frequency of African Easterly waves (Bain et al. (2014)), and their link to mesoscale convective systems (Tomassini (2018)). At the 21 21 heart of the difficulty in representing these phenomena (in the current and earlier generations of GCM resolutions), lies the need to enable appropriate positive feedbacks to develop between convection and its parent dynamics, while keeping the model stable by preventing the emergence of convective instability onto the resolved grid. 

26 25 The need for model stability has similarly made an apparently straightforward problem of capturing the correct timing (and amplitude) of the diurnal cycle of convection over land (Yang and Slingo (2001), Christopoulos and Schnei-der (2021), and Tao and et al (2024)) a challenge to rectify, as delaying convective activity for too long in the presence of a convectively-unstable atmosphere risks instability reaching the scale of the model grid, and the resolved dynamics 30 28 responding accordingly. 

Mass-flux schemes that use a closure based on convective available potential energy (CAPE) have success in keeping the resolved model stable (e.g. Walters et al. (2019)), but the use of such a vertical integral of convective 34 31 instability in the troposphere to determine the mass flux at cloud base produces a disconnection between the condi-tions required in the boundary layer to trigger convection (for example, enough kinetic energy to overcome layers of convective inhibition), and those linked to its response (e.g. Mapes (2000) and Fletcher and Bretherton (2010). This can lead to numerical intermittency of CAPE-closed schemes across the timestep, as the CAPE-closure can create 40 36 enough convective inhibition to prevent triggering on the next timestep (e.g. Klingaman et al. (2017)). Furthermore closures based on CAPE desensitise convective activity to local variations in the vertical thermodynamic profile, mak-ing it harder for the convection scheme to respond in tandem with the resolved dynamics (Whitall et al. (2022)). In addition, the use of predefined profiles for entrainment leads to somewhat inflexible representations of the mass-flux 45 40 shape, making smooth transitions between different convective states hard to achieve.

These limitations, along with embedded decisions in the convection-scheme code about the levels of convective initiation, and the number of updraughts allowed at any one level, have motivated the development of an entirely new 

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convection scheme called CoMorph, Whitall et al. (2022). This scheme provides a much more flexible architecture
that enables complexity to be built in as required, and with the aim of creating a scheme that is better connected to
the underpinning physical processes that drive convective activity.

CoMorph-A represents a first implementation of this scheme, designed for a global package, with the priorities of this version being to improve the coupling between convection and the resolved dynamics, as well as the consistency of interaction with other physics parametrisations. Numerous tests of the scheme under controlled forcing conditions have been conducted, for example in a Single Column Model (Whitall et al. (2022)), and for a range of case studies in the Idealised MetUM (a 3D, Cartesian, bi-periodic version of the Met Office Unified Model, Lavender et al. (2024)), with promising behaviour across a range of regimes. The coupling of the scheme to the dynamics has also been investigated (Daleu et al. (2023)), again highlighting the significant advances and potential of the new scheme. Zhu et al. (2023b) have also investigated the impacts of CoMorph-A in global simulations, focusing on the Indo-Pacific and Australian regions, finding significant benefits. 

This paper introduces the CoMorph-A and accompanying control package configurations of the Met Office Uni-fied Model (UM) in sections 2 and 3. To evaluate the impacts and performance of the new package, we follow the standard testing procedure as laid out in Walters et al. (2019), and evaluate the performance of the same configura-tion in both climate and numerical weather prediction (NWP). The climate evaluation in section 4 uses 20 years of simulation using AMIP specified sea-surface temperatures and sea ice. For NWP, while we have also run forecast only 22 59 simulations, as in Walters et al. (2019), in section 5 we present verification from two 3-month long cycling forecast and analysis trials of the CoMorph-A package, so including 4DVar data assimilation (albeit without updating the error covariance and other assimilation statistics). 

# 29 63 2 | CONTROL MODEL

The control configuration of the Met Office Unified Model (MetUM) used in this study is GA8GL9, which is based on GA7GL7 (Walters et al. (2019)) but incorporates new developments intended for the next release that will be documented in detail elsewhere. GA8GL9 is the basis for the current operational NWP at the Met Office at the time of writing. The main changes since GA7GL7 are:

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   38 a package of changes related to model drag, described in Williams et al. (2020)
- revised parametrizations in the grid-scale microphysics for riming and the depositional growth of ice (Furtado and
   Field (2017))
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  71 a new "multigrid" dynamical solver
- reduced drag at high wind speeds over the sea which improves forecasts of tropical cyclones at higher model
   resolutions than discussed here
- 45 74 significant modifications to the existing massflux convection scheme 46
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   48 75 To give a brief description of GA7GL7, the ENDGame dynamical core uses a semi-implicit semi-Lagrangian formulation

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1 2 76 to solve the non-hydrostatic, fully compressible deep-atmosphere equations of motion (Wood et al. (1999)). The 77 grid-scale microphysics scheme (i.e., that which operates on the gridded prognostic variables) is single-moment with prognostic rain and ice, based on Wilson and Ballard (1999). The parametrisation of the fraction of the grid box 78 which is covered by cloud, and the amount and phase of condensed cloud water it contains, is the prognostic cloud 79 fraction and prognostic condensate (PC2) scheme (Wilson et al. (2008a), Wilson et al. (2008b)). The parametrization 80 81 of radiative transfer (solar, shortwave (SW), and thermal, longwave (LW)) uses the SOCRATES scheme (Edwards and Slingo (1996); Manners et al. (2015) The turbulence scheme uses a first-order closure from Lock et al. (2000), mixing 82 adiabatically conserved heat and moisture variables, momentum and tracers, using non-local profiles of diffusion 83 coefficients within convective boundary layers and an explicit parametrization of entrainment across their top. In the 84 85 control simulations the parametrization of sub-grid-scale transport of heat, moisture and momentum associated with cumulus clouds is a mass-flux convection scheme based on Gregory and Rowntree (1990). Three separate classes of 86 convection are represented. First a diagnosis is made to determine whether convection is possible from the boundary 87 layer, and its potential depth, using an adiabatic undilute parcel ascent, which then triggers either a shallow or deep 88 convection scheme. Any remaining instability to moist ascent identified above the boundary layer and the top of any 80 shallow or deep convection is handled by a mid-level scheme. Each component involves different parametrization 90 of processes such as entrainment and detrainment from the plume, together with separate closures (based on the 91 surface buoyancy flux for shallow convection and CAPE for mid-level and deep). 22 92

Of the modifications to the GA7GL7 convection scheme used in our control, two are of most relevance here. The 93 24 first, described in Willett and Whitall (2017), is to relate the entrainment rate in the deep component of the scheme to 94 25 the amount of recent convective activity, as measured by an advected prognostic, similar to Mapes and Neale (2011). 95 26 27 96 The source term for this is scaled on the surface convective precipitation rate, expanded to three dimensions using 28 the current convective temperature increment profile, and decays with a 3 hour timescale. In this way the scheme 97 29 is given a short term memory of previous convective activity, thereby allowing evolution of convective clouds over 98 30 finite time scales. The second significant change from GA7GL7 is to time-smooth the heat and moisture increments 99 31 32 100 from the convection scheme over a timescale of 45 minutes. This is to reduce the impact of strong time-step-level 33 101 intermittency in the scheme (see section 4) on the resolved scale dynamics. Note that this increment smoothing is 34 not used with the CoMorph-A package. 102 35

Finally, note that most of these parametrizations have undergone significant modification between their original 36 103 37 104 references, as above, and GA7GL7 and the reader is referred to Walters et al. (2019) for more detail.

#### 40 105 **COMORPH-A PACKAGE** 3

An important aspect of the new convection parametrization, CoMorph, is that it is directly coupled to, and hence 43 106 44 107 strongly dependent on, other aspects of the model physics. In particular, the presence or otherwise of cloud strongly 45 108 affects the initial sources of mass, and the turbulence statistics from the boundary layer scheme determine the ini-46 tial characteristics of the plume. Initial development in a single-column version of the MetUM highlighted unwanted 109 47 sources of intermittency and sporadic unrealistic behaviour at the time step and grid level. For the proper functioning 48 110

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111 of the CoMorph convection scheme, as well as for the overall performance, it was therefore necessary to make signif-112 icant changes to several other aspects of the physical parametrizations that will be described in this section. Together 113 these then form the CoMorph-A package.

#### 3.1 CoMorph convection scheme 114

The control convection scheme is replaced by the entirely new scheme, CoMorph. Details are given in Whitall et al. 115 (2022) but we give a summary here. 11 116

12 117 As in the control, the CoMorph scheme uses the massflux approach to parametrize the effects of subgrid cumulus 13 118 convection but CoMorph has an entirely new code structure, to allow greater flexibility, and many differences in 14 approach with additional physical processes included. There is no external closure in CoMorph. Instead, the massflux 15 119 initiated from any locally unstable model grid-level is proportional to the relative size of the local environment lapse 16 120 17 121 rate to that of a test parcel. Where the prognostic cloud fraction parametrization identifies a grid box as having partial 18 122 cloud cover, separate initiation calculations are made in the saturated and unsaturated air and these are combined 19 using area weighting by the cloud fraction. 20 123

The rate of entrainment scales with an inverse parcel radius, i.e., 21 124

$$\epsilon = \frac{\alpha_{ent}}{R} \tag{1}$$

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26 where  $\alpha_{ent}$  is a dimensionless parameter currently set to 0.2. The parcel radius, R, is based on a turbulence length-125 27 scale in the parcel's source-layer given by  $R = a_L K_M / \sigma_w$ , where  $K_M$  is the momentum diffusivity in the boundary 28 126 layer scheme and  $a_L$  a constant taken to be 0.45;  $\sigma_w = (K_M/\tau)^{0.5}$  is a turbulent velocity scale where  $\tau$  is a turbulent 29 127 30 128 timescale, see Van Weverberg et al. (2016). Initial testing in the global model, however, found this simple formulation 31 was insufficient to remove instability fast enough under all circumstances and so an additional scaling was introduced, 129 32 increasing  $a_L$  based on an ad-hoc linear function of the previous time step's precipitation rate (up to a factor of 33 130 2 for a precipitation rate above approximately 35 mm per day). Qualitatively this can be viewed as allowing the 34 131 35 132 updraft size to respond to the organisation of convection via precipitation leading to the formation of cold pools and 36 subsequent initiation of larger-scale updraughts (Mapes and Neale (2011)). The parcel radius increases with height 133 37 through entrainment by assuming the mass-flux is carried by a fixed number-flux of spherical thermals whose radii 38 <sup>134</sup> 39 135 increase consistent with the increase in volume of each thermal.

40 136 The detrainment rate is modelled as removing the non-buoyant part of an assumed distribution of buoyancy 41 within the bulk plume, informed by separate parcel ascent calculations for parcel mean and less dilute core properties. 137 42 Importantly, this detrainment calculation is solved implicitly, in which the convective heating of the environment (as 43 138 44 139 well as the other model increments through the timestep) is accounted for in the calculation of the parcels' buoyancy 45 140 and therefore the proportion of the buoyancy distribution that is detrained.

46 The initiating parcel mean properties are calculated separately within each clear or cloudy sub-grid region, as with 141 47 massflux initiation, and then averaged together weighted by their respective initiating massfluxes. The in-region prop-48 142

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143 erties are taken as just saturated and neutrally buoyant and then given a perturbation,  $\phi'$ , linked to the parametrized turbulent flux ( $\phi' = \alpha_t \overline{w' \phi'} / \sigma_w$  with  $\alpha_t = 1/3$  and  $\phi$  being temperature, water vapour mixing ratio and the horizon-144 tal wind components). Because the parcel core is representing the most extreme tail of the distribution of sub-grid 145

fluctuations, the core property perturbations are scaled up by a factor of 3.

As with the control, CoMorph uses a massflux representation of momentum transport by convection that also 147 includes the effects of cloud pressure-gradients on the flow within the cloud (e.g., Kershaw and Cregory (1997)). 148

Importantly, CoMorph also includes a novel in-parcel bulk microphysics scheme for the generation of hydromete-10 149 11 150 ors. These hydrometeors are then passed on the model-level where they leave the parcel (through detrainment and 151 a fall-out flux) to the grid-scale microphysical variables. This can be achieved because, as in the control, the model's grid-scale (Wilson-Ballard) microphysics scheme prognoses the mass concentration of each hydrometeor species so 152 convection can now provide an additional source term for these prognostics. Their fall to the surface, and any inter-15 153 16 154 actions on the way down, is modelled by the Wilson-Ballard scheme which therefore improves the consistency of 17 155 treatment of precipitation between that generated through convective processes and that through larger, resolved scale ones. Note that diagnostically this means there is no longer any separation into convective or large-scale pre-156 cipitation, just precipitation. In order for convection to modify rain mass and area fraction consistently a prognostic 20 157 21 158 precipitation area fraction is introduced and passed to the large-scale microphysics. In addition to cloud liquid water, 22 159 ice and rain, CoMorph itself includes graupel, which is not represented explicitly in the control model, being largely a 160 feature of deep convection which is entirely parametrized there, but passing CoMorph's graupel to the large-scale's single ice variable led to far too excessive build-up of cloud ice in the tropics and so the large-scale microphysics' prog-25 161 26 162 nostic graupel was also included in the CoMorph-A package (following the configuration already used in the regional 27 163 configurations of the UM, see Bush et al. (2020)). This initial implementation of a microphysics scheme in CoMorph 164 is relatively simple, prescribing a representative number concentration for each hydrometeor species in order to cal-30 165 culate an effective particle size and thence fall speeds, accretion rates and other exchanges between hydrometeor 31 166 classes. Crucially this allows the parcel (and detrained air) to become supersaturated with respect to ice, whereas the 32 167 convection scheme in the control adjusts the parcel fully to ice-saturation when below the melting point temperature.

33 CoMorph is also written in terms of mixing ratios of water vapour and condensate (while the control convection 34 <sup>168</sup> 35 169 scheme uses specific quantities). This has then allowed all the model's physics schemes to move across to using mixing 36 170 ratios, consistent with the dynamical core, and so avoid conversions in moving between different parts of the model 37 <sub>171</sub> which allows global water conservation errors to be reduced to negligible levels. 38

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- it is a single unified scheme for all convective regimes
- it is closely integrated with the rest of the model physics, giving greater physical consistency 43 174

In summary, the key aspects of the CoMorph convection scheme are:

- 44 175 massflux is initiated through local moist instability leading to a much tighter coupling with the resolved dynamics 45 176 of the model
- 46 close attention has been paid to the numerical methods used, with substantial aspects solved implicitly, and this 177 47 leads to a smooth temporal behaviour 48 178
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2 179	it includes a microphysical representation of cloud process in its updraughts and precipitation is passed directly
3 <sub>180</sub> 4	to the prognostic grid-scale microphysical variables instead of being rained out diagnostically, thereby improving
- <sub>181</sub>	consistency of treatment of microphysical processes in the model
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7 0 <sup>182</sup>	3.2   Other model changes
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10 183	3.2.1   Cloud scheme
11 184	The impact of triggering CoMorph from either saturated or unsaturated air makes it sensitive to the diagnosis of the
12	fraction of the grid box that is cloudy. In the process of developing CoMorph, several aspects of the PC2 cloud scheme
13 14 186	were discovered to be excessively sensitive to small perturbations. Largely motivated, then, by a desire to reduce the
15 187	timestep level variability of convection, the following changes were made to the PC2 cloud scheme:
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17 10 <sup>188</sup>	• to ensure the cloud is consistent with the thermodynamic profiles at the point in the timestep where these are
10 19 189	passed to CoMorph, the PC2 homogeneous forcing calculation from advection is moved from the end of the
20 190	timestep to just after the advection calculation (which is called before convection) and new calls to the PC2
21	initiation and checking algorithms are added before the call to CoMorph
22 23 <sup>192</sup>	• the "BiModal" scheme of Weverberg et al. (2021) is used for initiation of cloud within PC2 (because of its more
24 193	realistic treatment of subgrid variability), together with using that initiated cloud as a minimum cloudiness (cloud
25 <sub>194</sub>	fraction and liquid water content) at every grid-point on every timestep (in the control the cloud has to dissipate
26 27 <sup>195</sup>	completely before initiation is triggered which can lead to sudden large changes in cloud fraction)
27 28 <sup>196</sup>	• for consistency, the pressure change experienced by the environment as it undergoes convectively forced subsi-
29 197	dence is used to drive the homogeneous forcing of cloud, following the same method as for resolved advection
30 198	(Wilson et al. (2008a)), a process not treated explicitly in the control
31 22	• an implicit numerical method is introduced to calculate the PC2 term for the erosion of cloud, which depends on
33 <sup>200</sup>	the cloud fraction (see Morcrette (2012)), such that the erosion rate applied is consistent with the cloud-fraction
34 201	that will occur after the erosion increment has been added on
35 202	• the method of Morcrette (2020) is used to ensure that the PC2 cloud scheme produces reasonable magnitude
36 37	cloud fraction increments in the limit of small condensate amounts
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39	322 L Cloud microphysics
40 204	
41 42 <sup>205</sup>	As discussed in section 3.1, the main changes to microphysics are to include graupel as a prognostic variable, which
43 206	allows for the explicit representation of a second, more dense ice category, and to include a fractional area of precipita-
44 207	tion so as to discriminate convectively generated rain that occupies a small fraction of the grid box from that generated
45 <sub>208</sub> 46	by large-scale processes. Another, more minor, improvement is to relax the shallow atmosphere assumption in the
40 47	calculation of flux divergences, and so be consistent with the rest of the model, which improves local conservation of

- 48 210 water.

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#### 3.2.3 **Fountain Buster** 211

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212 In the absence of sufficient lateral subgrid mixing, the MetUM has been found susceptible to the formation of neargrid-scale flow structures that, combined with the lack of inherent conservation with semi-implicit semi-Lagrangian 213 advection, leads to significant conservation errors. This problem can be illustrated by considering an idealised up-214 draught in a single grid-column with continuity requiring a convergent in-flow in lower levels. The interpolation of 215 the Arakawa C staggered horizontal wind fields onto scalar points, however, strongly reduces resolved lateral conver-10 216 11 217 gence from surrounding points into grid-scale vertically ascending plumes. As a result, the vertical transport of, for 12 218 example, water vapour away from the near surface is not compensated for by horizontal convergence of drier air into 13 14 219 the plume from the sides, with the result that moisture can be transported vertically ad infinitum. Typically this is then 15 220 manifested as extreme local precipitation accumulations. The use of a posteriori global conservation correction has 16 221 been found to reduce these errors significantly (Bush et al. (2020)), but substantial local errors have been found to 17 remain. 222 18

Here a post-hoc local correction is applied after the semi-Lagrangian scheme's interpolation to the departure 20 223 21 224 point, in the form of the linear up-wind advection increments that arise from the convergent part of the flow that 22 225 is removed when the horizontal wind components are interpolated to the scalar grid points for the departure point 23 24 226 calculation. In more detail, first, at each lateral cell face a measure of the (assumed missing) convergent stagnation in-flow, S, is calculated. For example, for the west cell face of scalar point (i, j, k): 25 227

$$S_{\text{west}} = \frac{u_{i-1,j,k} - s_d u^{\theta}}{u_{i-1,j,k}}$$
(2)

30 where  $u^{\theta}$  is a linear estimate of u at the scalar point. To ensure only the convergent part of the flow is accounted for, 31 228 32 229 we ensure  $0 < S_{west} < 1$ , and to avoid making increments in reasonably well-resolved flows the tuning factor,  $s_d$ , is 33 230 set to 2. 34

The fountain buster tendency for a scalar variable  $\chi$  is then given by first-order upwind advection scaled by S, 231 37 232 i.e.:

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 $\begin{array}{lll} \frac{\Delta\chi_{i,j,k}}{\Delta t} & = & S_{\mathrm{east}}u_{i,j,k}^{f}\frac{\chi_{i+1,j,k}-\chi_{i,j,k}}{\Delta x} - S_{\mathrm{west}}u_{i-1,j,k}^{f}\frac{\chi_{i,j,k}-\chi_{i-1,j,k}}{\Delta x} \\ & + & S_{\mathrm{south}}v_{i,j-1,k}^{f}\frac{\chi_{i,j,k}-\chi_{i,j-1,k}}{\Delta y} - S_{\mathrm{north}}v_{i,j,k}^{f}\frac{\chi_{i,j+1,k}-\chi_{i,j,k}}{\Delta y} \end{array}$ (3)

44 233 where the superscript f denotes the wind interpolated (vertically) to the face centre of the scalar point. In (3) a regular 45 234 cartesian grid has been used for simplicity but in the MetUM the appropriate metric variables are used. The fountain 46 buster is applied to all advected scalar variables, so all moisture variables, tracers and the thermodynamic variable (dry 235 47 virtual potential temperature). 48 236

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#### 3.2.4 **Turbulent mixing** 237

A deliberate choice in the Lock et al. (2000) boundary layer scheme was to interface with the convection scheme 4 238 5 at the lifting condensation level (LCL, diagnosed from the top of the surface layer), meaning that the boundary layer 239 6 scheme would be entirely responsible for surface-driven mixing at least up to the LCL, from where the convection 240 7 scheme would then be initiated and take over the transport into the free troposphere. With CoMorph, initiation can 241 8 now occur at any level and so we take the opportunity to relax the requirement that the boundary layer scheme mix 9 242 10 243 up to this definition of the LCL. We still use the Lock et al. (2000) undilute moist adiabatic parcel ascent method to 11 244 diagnose the potential for cumulus convection but, in those regimes, now diagnose the top of the surface-driven 12 turbulence diffusion profile as the point where the integral of diagnosed negative buoyancy flux is a fraction (0.05, 245 13 see Walters et al. (2019)) of the vertical integral of the positive flux. Because we assume any transport within clouds 14 246 15 247 will be carried by CoMorph, we use the unsaturated buoyancy flux in this calculation.

16 248 A second alteration to the boundary layer scheme is in the calculation of the buoyancy gradient used in the 17 18 <sup>249</sup> Richardson number (Ri). This is revised to reduce the mixing from the Ri-dependent part of the scheme across statically stable inversions, in particular where these cap a cloudy boundary layer. With the UM's Charney-Philips 19 250 20 251 vertical grid staggering,  $R_i$  is required on the same grid-level as scalars (referred to as  $\theta$ -levels), where the vertical 21 gradient of horizontal momentum naturally falls. In the control, the gradients of liquid-ice static energy temperature 252 22 and total water content are interpolated to  $\theta$ -levels and that  $\theta$ -level's cloud fraction used to obtain the buoyancy 23 253 gradient. It is found that at cloudy  $\theta$ -levels below a strong inversion this interpolation of the gradients, combined 24 254 25 255 with a saturated buoyancy calculation, can lead to apparent instability. In the CoMorph-A package we calculate the 26 256 buoyancy gradient locally (on  $\rho$ -levels) and interpolate that to the  $\theta$ -levels. To calculate the saturated contribution to 27 the buoyancy gradient on  $\rho$ -levels, an estimate of the vertical fraction of the grid containing saturated air is made from 28 257 interpolating the supersaturation in the adjacent  $\theta$ -levels. In this way the contribution from the  $\rho$ -level with strong 29 258 30 259 gradients is typically largely unsaturated, leading to strong stability, which then remains stable when averaged with 31 260 the saturated  $\rho$ -level with weak gradients below. 32

As with the Fountain Buster, the more active dynamics in the tropics arising from more organised convection with 33 <sup>261</sup> 34 262 CoMorph also motivates enabling the Leonard terms (Hanley et al. (2019)), which include extra subgrid vertical fluxes 35 263 that account for the tilting of horizontal flux into the vertical by horizontal gradients in vertical velocity. Including these 36 264 terms helps to weaken a few occurrences of excessively strong resolved updraughts and precipitation in organised 37 convective structures. 38 <sup>265</sup>

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#### Tuning of the top-of-atmosphere radiation 3.3

As with any major model developments, the initial simulations showed significantly worse biases than the control, 43 267 44 268 both locally and globally in the top-of-atmosphere radiation. To reduce excessively weak outgoing longwave radiation 45 269 (OLR) in the tropics, we increased the fall speeds for cloud ice by 20% (which increases the global mean OLR by around 46 1.5 Wm<sup>-1</sup> and reduces the spatial root-mean-square error (rmse) against satellite climatology (see section 4) by 0.3 270 47 Wm<sup>-1</sup>) and also removed a parametrization in PC2 that increased the ice cloud fraction through vertical wind shear 48 271

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272 that was not especially well justified physically (which increased the global mean OLR, by 0.5 Wm<sup>-1</sup>, and reduced the rmse by  $0.3 \text{ Wm}^{-1}$ ). In the shortwave (SW), the cloud forcing was initially too strong across most of the tropics. 273 Several options are available that can reduce the reflectivity of clouds and we have made two revisions. The first was 274 to the parameters in the Liu et al. (2008) spectral dispersion of the cloud droplet size distribution. As described in 275 Mulcahy et al. (2018), the cloud droplet spectral dispersion as implemented in the control is given by 276

$$\beta = a \left(\frac{L}{N_d}\right)^b \tag{4}$$

where L is the cloud liquid water content and  $N_d$  the cloud droplet number concentration. In the control a = 0.07 and 277 b = -0.14 while in the CoMorph-A package we use a = 0.093 and b = -0.13. These changes, that are small compared 14 278 15 279 to the spread in the Liu et al. (2008) data, resulted in a reduction in global mean reflected SW of around 1.5 Wm<sup>-1</sup>, 16 280 whilst having almost no impact on OLR. Regionally the SW impact was strongest broadly across the marine sub-tropics 18 <sup>281</sup> and gave a reduction in spatial rmse of around 1 Wm<sup>-1</sup>. Our second tuning, to further reduce the reflectivity of clouds, was to increase a parameter in the fractional standard deviation of liquid clouds (equal to the standard deviation of 19 282 20 283 cloud water content in a grid box divided by its mean value, see Hill et al. (2015)) from 1.6 to 1.65, which reduced the 21 284 global mean reflected SW by around 0.7 Wm<sup>-1</sup> and the spatial rmse by 0.25 Wm<sup>-1</sup>. Note, though, that the impacts of these last two changes would certainly be different if implemented in the reverse order. 285

24 286 It is important to remember that this tuning is a critical step in any new physics package and was, of course, also 25 287 done for the control configuration. Hence the impacts discussed in section 4 should be viewed as resulting from the CoMorph-A package as a whole. 288

#### 3.4 Cost 30 289

There is a marginal increase in the CPU time with CoMorph-A of roughly 5% over the control of which a substantial 32 290 33 291 part will be the new prognostic graupel - tests with just the convection scheme reverted to that in the control show 34 <sub>292</sub> a similar increase indicating the CoMorph convection scheme itself is cost neutral.

#### **RESULTS IN AMIP** 4 38 <sup>293</sup>

40 294 We evaluate the climatology of the CoMorph-A package using AMIP simulations at N96 resolution (around 135 km 41 42 295 grid spacing in mid-latitudes). Given the main change is to the convection parametrization, we start by looking in Fig. 1 at the impact on the annual mean precipitation. The most systematic change is to have increased rain over tropical 43 296 44 297 land, including the maritime continent, most of which is beneficial compared to the Global Precipitation Climatology 45 298 Project (GPCP, Adler et al. (2003)). Particularly relevant for the Maritime Continent, a significant issue noted by opera-46 tional meteorologists with the control simulation in the tropics is precipitation in convective weather systems abruptly 299 47 stopping at the coastline as they move from sea to land. This issue is much improved in CoMorph-A. Especially over 48 300

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301 the Pacific Ocean there is a sharpening of the ITCZ, giving enhanced rainfall on the equator with reductions to the 302 north and south, with mixed impact.

4 More detailed analysis shows the characteristics of the model's tropical precipitation at local time and space scales 303 5 are altered beyond recognition. Fig. 2 quantifies the probability of grid point precipitation rates at each timestep 6 304 7 given the rate at the previous timestep. In the control model there is a strong signal along the axes, indicating a 305 8 high probability of no precipitation on one timestep if there was precipitation on the previous, and vice versa - this 306 9 10 <sup>307</sup> illustrates the long-standing, Klingaman et al. (2017), time-step level intermittency in the control. With CoMorph-A this timestep intermittency is completely removed and shows a far more plausible temporal coherence with high 11 308 12 309 probabilities around the one-to-one line. This continuity between time steps is a direct result of the use of improved 13 310 numerical methods, as discussed in section 3, and the implicit solution for detrainment in CoMorph, in particular, as 14 this will tend to maintain similar convective inhibition into the next time step. Interestingly, after averaging over 3 15 311 hours and 2x2 grid points for comparison with the GPM observations (Global Precipitation Measurement (GPM) Multi-16 312 17 313 satellitE Retrievals (IMERG) precipitation data V06B, Tan et al. (2019)), the models' precipitation characteristics are far 18 314 more similar. This convergence is similar to that seen for a range of MetUM configurations at various resolutions in 19 Martin et al. (2017) and similar to the results for a number of CMIP5 models (except for those with particularly coarse 20 315 resolution) in Klingaman et al. (2017). There is, nevertheless, a suggestion of more temporally coherent heavy rain 21 316 22 317 with CoMorph-A, closer to GPM, albeit still insufficient. It is also interesting to reflect that many of the geographical 23 318 biases in tropical precipitation seen in Fig. 1 are remarkably similar, despite entirely replacing the convection scheme, 24 25 <sup>319</sup> although many of these simply reflect the regions with the heaviest rainfall.

26 320 The strength of propagating tropical waves is strongly improved, see Fig. 3, both for Kelvin waves (as marked 27 <sub>321</sub> on the figure) and the Madden-Julian Oscillation (MJO, the region of strong observed eastward-propagating 30-60 28 29 <sup>322</sup> day variability at wave numbers 1 to 3). We believe this tighter coupling between CoMorph and the resolved dynamics arises because the initiation of massflux can respond to locally generated moist instability, but more detailed 30 323 31 324 investigation will be the subject of future work.

32 Fig 4 shows a large increase in the integrated tropical water vapour that all but removes a dry bias in the control, 325 33 largely occurring in the upper troposphere. Around a half of that moistening comes from the Fountain Buster scheme, 34 326 35 327 while the rest is likely due to passing the precipitation into the "large-scale" microphysics that then represents its 36 <sub>328</sub> evaporation as it falls to the surface more realistically (although to confirm this would require adding some form of 38 <sup>329</sup> direct precipitation to the surface within CoMorph, which is beyond the scope of this work). This moister troposphere 39 330 may also contribute to the improvement in the strength of tropical waves, see Fig. 3 (Zhu et al. (2023a)).

40 331 There are also substantial changes to the clouds in the model, as seen in the top-of-atmosphere cloud radiative 41 42 332 effect (CRE, which is the clear-sky minus the total radiative flux). The longwave CRE, Fig. 5, is much stronger across the tropics and especially over tropical land, which all but eliminates a serious underestimate in the control. This is due to 43 333 44 334 a large increase in the tropical ice cloud amount (both mass and cloud fraction) as can also be seen in the comparison 45 335 with CALIPSO observations over central Africa in Fig. 6, likely arising from the handling of precipitation sedimentation 46 by the grid-scale microphysics. However, like the control, there is less cloud than observed by CALIPSO between 5km 336 47 and 8km altitude suggesting that the congestus phase remains poorly simulated. The increase in optically thick cirrus 48 337

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338 means shortwave CRE (Fig. 7) is also enhanced somewhat over tropical land, but the biggest impacts here are in the subtropics and especially over the eastern Pacific and Atlantic. Climatalogically, these are areas dominated by stratocu-339 mulus clouds. The geographical pattern of the change from including the CoMorph-A package matches remarkably 340 well with the error in the control, although the end result is for the clouds to be somewhat too reflective over most of 341 the tropical oceans. Part of the increase in stratocumulus arises from the implementation of BiModal initiation in the 342 343 PC2 cloud scheme and the revised calculation of Ri, but around half comes from CoMorph itself, especially further west from the coastline where shallow cumulus clouds form under the lifting stratocumulus. Detailed single column 344 10 model analysis of this transition region (not shown) indicates that the combination of the smoother temporal evolution 345 11 of CoMorph (as in Fig 2) and more sensitive (implicit) detrainment allow a smoother evolution of the stratiform cloud 12 346 13 347 layer as it rises away from the coast. Consistent with the improved top-of-atmosphere radiative fluxes, we also find 14 the net surface energy fluxes (that are important for coupled ocean modelling) are significantly improved, and across 348 15 the tropics and subtropics especially, compared to the observed estimate from Liu et al. (2015) (not shown). 349 16

17 350 A form of convective "memory" was implemented in the control model (in GA8GL9 relative to GA7GL7, as de-18 351 scribed in section 2) to delay the development of deep convection and help address the poor representation of the 19 diurnal cycle of precipitation over land. However, consistent with other studies (Willett and Whitall (2017); Tao and 352 20 21 353 et al (2024)), we find GA8GL9 still has issues in simulating the correct timing of the diurnal cycle over many areas of 22 354 the globe (see Fig. 8). CoMorph-A does not have such a memory function and has a tendency to have the diurnal peak 23 355 in rainfall intensity over tropical land around local noon or early afternoon, compared to the observed late evening 24 356 peak (Fig. 8). Over some regions (e.g. Africa) this is earlier and hence detrimental compared with the control, whilst 25 over others (e.g. S. E. Asia, as also seen in Zhu et al. (2023b)) CoMorph-A is an improvement on the control, although 26 357 27 358 still considerably earlier than observed. Diurnal cycle experiments using the idealised MetUM (Lavender et al. (2024)) 28 359 also show too early initiation and development to deep convection using CoMorph-A compared to high resolution 29 360 simulations, although improved relative to the control. 30

#### 5 **RESULTS IN NWP**

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35 <sub>362</sub> The NWP trials (cycling forecasts and data assimilation) are performed at somewhat higher, N320 (~ 40 km in mid-36 37 <sup>363</sup> latitudes), resolution than the climate simulations presented in section 4. A wide variety of metrics are monitored and these are summarised in the "scorecards" shown in Fig. 9. These illustrate the change in root-mean-square error (rmse) 38 364 39 365 against each trial's own analysis (scorecards are also produced against observations and independent global analyses 40 366 from ECMWF but those look very similar so are not shown). The majority of metrics show reductions in rmse with 41 42 <sup>367</sup> CoMorph-A, with somewhat better overall performance in the boreal summer than winter (overall reductions of 1.9% and 0.5% respectively). Some degradation is seen in a number of fields in the first two days that may arise from not 43 <sup>368</sup> 44 369 updating the error covariance statistics in the trials. One field that shows degradation throughout the forecasts in both 45 370 summer and winter, though, is the tropical temperature at 850 hPa. Fig. 10 shows the geographical distribution of the 46 change in rmse at day 6, which has large positive values across most of the tropical oceans but especially so, for these 371 47 austral summer trials, in the sub-tropical south-eastern Pacific and Atlantic close to the edges of where CoMorph-A 48 372

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2 373 gives more stratocumulus cloud layers (the NWP trials show a similar increase to that shown to be beneficial in the 3 374 AMIP simulations in Fig. 7). Fig. 10 also shows that the standard deviation of temperature at 850 hPa in the analyses is 4 375 increased in CoMorph-A, and the day 6 forecasts in CoMorph-A have similarly more variable temperatures at this level 5 (not shown). The more persistent stratocumulus clouds with CoMorph-A will result in a sharper temperature inversion 376 6 7 at the top of the boundary layer. Furthermore, on the western edge (and further downwind) of the stratocumulus, 377 8 378 there is more variability of the cloud cover as it transitions to shallow cumulus. These transitions result in variations in 9 the height of the inversion which will be manifested as increased temperature variability and rmse at this level, even 379 10 if the average inversion height or boundary layer temperature were accurately forecast (the 1000 hPa temperature 380 11 rmse is actually improved or neutral over tropical oceans, not shown). 12 381

13 382 A long standing systematic error is a slow bias in the winter extra-tropical jet. Fig. 11 shows an increase in the wind 14 383 speed from CoMorph-A, reducing this error. It is often the case that increasing the wind speed will also increase the 15 384 rmse due to a "double penalty" if a wind feature is geographically displaced. It is notable, therefore, that CoMorph-A 16 reduces the slow bias with an overall neutral impact on rmse for upper level winds (Fig. 11 & Fig. 9). 17 385

18 386 Further analysis of the trials has included tracking the location of extra-tropical cyclones through the maximum 19 in 850 hPa vorticity (Hodges (1995)). Whilst the error bars overlap and hence cannot be regarded as significantly 387 20 21 388 different, the errors in these tracks are reduced in CoMorph-A and experience suggests the fact the signal is consistent 22 389 in the two seasons is notable (Fig. 12).

23 390 Finally, between 25 June to 31 October 2021, atmosphere-only forecasts with the CoMorph-A package were run 24 daily at the operational resolution of N1280 from Met Office operational analyses at 0 UTC. Verification of tropical 391 25 cyclones from these forecasts in Fig. 13 shows CoMorph-A gave substantial deepening and improvement to wind 26 392 27 393 speeds. Whilst the central pressure in these simulations is now lower than observed, they are atmosphere-only and 28 394 we would expect slower-moving storms to be weaker when coupled to an ocean model in operational systems. 29

#### CONCLUSIONS 6

34 396 The results from a range of global simulations have been presented for a large package of revised parametrization 35 <sub>397</sub> schemes, centred around the entirely new CoMorph massflux convection scheme. The key components of CoMorph are that it has a physically-based flexible formulation that allows it to represent all convective regimes and is tightly 398 37 coupled to the rest of the model, both the other parametrization schemes and the dynamics. The main impacts of this 38 <sup>399</sup> 39 400 overall CoMorph-A package in standard global test simulations are found to be:

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- removal of the timestep level intermittency of convection that has been in all configurations of the MetUM since 401 42 43 402 its inception
- 44 403 improved tropospheric humidity over tropical land
- 45 404 improved tropical variability, including Kelvin waves and MJO
- 46 surface fluxes improved in many regions, important when coupling to an ocean model 405 47
- increased high cloud over tropical land where it has always been lacking and hence marks a significant improve-48 406
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- increased low cloud over tropical oceans which is generally better but optically too bright everywhere (although tuned for the global mean net top-of-atmosphere radiative flux, in order to balance weak clear sky out-going longwave radiation)
- improved NWP performance, including reduced tropical and extra-tropical cyclone errors
- some degradation of the diurnal cycle of tropical precipitation

11 413 Given the extent of new developments within the CoMorph-A package, this represents impressive performance for 12 414 a first implementation of a new convection scheme. Future work will focus on improving the diurnal cycle, through 13 415 for example inclusion of a second updraught to represent, separately, convection forced from cold pools, and on 14 CoMorph's applicability to higher resolutions, where the larger scales of convection begin to be resolved.

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# 37 430 Data availability statement

 $39_{431}$  The data that support the findings of this study are available from the corresponding author and Met Office co-authors  $40_{432}$  upon reasonable request.

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FIGURE 2 Histograms of the probability of bins of precipitation, measured at the time step and grid point level (top
 row) and over 3 hourly and 2x2 grid box averages (bottom row), aggregated over all grid points in the equatorial Indian
 Ocean from N96 AMIP simulations. The dashed lines show the one-dimensional histogram of the binned precipitation,
 using the right-hand vertical axis.



**FIGURE 4** Annual mean total column water vapour from N96 AMIP simulations, and compared to ERA40 (Uppala et al. (2005)) in the bottom row.

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**FIGURE 5** Longwave cloud radiative effect compared to Clouds and the Earth's Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) dataset (Loeb et al. (2009)) from N96 AMIP simulations



FIGURE 6 Histograms of height vs 532nm lidar backscatter ratio over central Africa (15°E-30°E, 20°S-10°N), showing climatologies of CALIPSO observations (Winker et al. (2002)) and simulated climatologies of CALIPSO data from 20-year N96 atmosphere/land-only climate simulations from the control and CoMorph-A







FIGURE 9 Scorecards against own analysis from the summer (left) and winter (right) NWP trials. Green upward triangles indicate reductions in root-mean-square error (rmse) and purple downward triangles increases, with the size of each scaled by the magnitude of change. Parameters are evaluated in the Northern Hemisphere (NH, north of 18.75° N), the tropics (TR, within 18.75° of the equator), the Southern Hemisphere (SH, south of 18.75° S), Europe (Euro) and UK (UK4 and UKIndex). The parameters are pressure at mean sea level (PMSL), vector wind (W), tempera-ture (T) and geopotential height (Z) at various pressure levels given in hectopascals (such that NH\_T250, for example, measures the change in rmse temperatures at 250 hPa in the northern hemisphere region). The columns represent forecast ranges from 6 h (T+6) to 7 days (T+168). 



**FIGURE 10** Change in root-mean-square error against own analysis of day 6 temperature at 850 hPa (top) and change in standard deviation of the analyses (bottom), from the winter NWP trials





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#### 1 2 3 4 ORIGINAL ARTICLE 1 5 **Journal Section** 6 7 8 The performance of the CoMorph-A convection 9 10 package in global simulations with the Met Office 11 12 Unified Model 13 14 15 16 A. P. Lock<sup>1</sup> | M. Whitall<sup>1</sup> | A. J. Stirling<sup>1</sup> | 17 K. D. Williams<sup>1</sup> | S. L. Lavender<sup>2,1</sup> C. Morcrette<sup>1</sup> 18 19 6 K. Matsubayashi<sup>3</sup> | P. R. Field<sup>1</sup> G. Martin<sup>1</sup> Т 20 21 M. Willett<sup>1</sup> | J. Heming<sup>1</sup> 22 23 24 <sup>1</sup>Met Office, FitzRoy Way, Exeter, EX1 3PB. 25 26 <sup>2</sup>Centre for Applied Climate Sciences, The impact on global simulations of a new package of physical 27 University of Southern Queensland, parametrizations in the Met Office Unified Model is documented. Toowoomba, Australia. 28 The main component of the package is an entirely new convection <sup>3</sup>Japan Meteorological Agency, Minato City, 29 scheme, CoMorph. This has a mass-flux structure that allows initi-Tokyo, Japan 30 ation of buoyant ascent from any level and the ability for plumes of differing originating levels to coexist in a grid box. it has a different 31 Correspondence form of closure, where the mass-flux of initiation is dependent on 32 Met Office, FitzRoy Way, Exeter, EX1 3PB. local instability, and an implicit numerical solution for detrainment 33 Email: adrian.lock@metoffice.gov.uk that yields smooth timestep behaviour. The scheme is more con-34 sistently coupled to the cloud, microphysics and boundary layer **Funding information** 35 parametrizations and, as a result, significant changes to these have also been made. The package, called CoMorph-A, has been tested 36 in a variety of single-column and idealised regimes. Here we test it 37 in global configurations and evaluate it against observations using 38 a range of standard metrics. Overall it is found to perform well 39 against the control. Biases in the climatologies of the radiative fluxes are significantly reduced across the tropics and sub-tropics. 40 tropical and extratropical cyclone statistics are improved and the 41 MJO and other propagating tropical waves are strengthened. It 42 also improves overall scores in NWP trials, without revisions to the 43 data assimilation. There is still work to do to improve the diurnal 44 cycle of precipitation over land, where the peak remains too close to the middle of the day. 45 46 KEYWORDS 47 Convection, parametrization, global, evaluation, Unified Model 48 49 1 50 51 52

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# 7 1 | INTRODUCTION

The representation of convection in general circulation models (GCMs) has long been recognised as having a central control on model performance, not only influencing the distribution of precipitation, but also global circulation pat-terns, and the resulting cloud and radiative fluxes that respond to them (e.g. Xie et al. (2012), Holloway et al. (2014) and Sherwood et al. (2014)). Common to many GCMs, the Unified Model has previously used a mass-flux scheme based on simplified Arakawa-Schubert principles (Arakawa and Schubert (1974)), which was devised by Gregory and Rowntree (1990), and used largely predefined profiles for entrainment, a simple link to detrainment, along with a CAPE-based closure. While this scheme and its subsequent variants (e.g. Derbyshire et al. (2011), Walters et al. (2019), Willett and Whitall (2017)) has been able to produce good mean atmospheric states (e.g. Walters et al. (2019)), it has had difficulty generating appropriate coupling between convection and the model dynamics. This is seen in a range of emergent atmospheric phenomena that rely on the close feedbacks between convection and the larger-scale dynam-ics. Examples, in the UM and many other GCMs, include the insufficient strength and progression of the MJO, (e.g. Klingaman and Woolhough (2013) and Ahn et al. (2020)); the low amplitude of convectively-coupled Kelvin and other equatorial waves (Vitart et al. (2007), Janiga et al. (2018) and Dias et al. (2023)); the limited strength and frequency of African Easterly waves (Bain et al. (2014)), and their link to mesoscale convective systems (Tomassini (2018)). At the 21 21 heart of the difficulty in representing these phenomena (in the current and earlier generations of GCM resolutions), lies the need to enable appropriate positive feedbacks to develop between convection and its parent dynamics, while keeping the model stable by preventing the emergence of convective instability onto the resolved grid. 

2625The need for model stability has similarly made an apparently straightforward problem of capturing the correct2726timing (and amplitude) of the diurnal cycle of convection over land (Yang and Slingo (2001), Christopoulos and Schnei-2827der (2021), and Tao and et al (2024)) a challenge to rectify, as delaying convective activity for too long in the presence3028of a convectively-unstable atmosphere risks instability reaching the scale of the model grid, and the resolved dynamics3129responding accordingly.

Mass-flux schemes that use a closure based on convective available potential energy (CAPE) have success in keeping the resolved model stable (e.g. Walters et al. (2019)), but the use of such a vertical integral of convective 34 31 instability in the troposphere to determine the mass flux at cloud base produces a disconnection between the condi-tions required in the boundary layer to trigger convection (for example, enough kinetic energy to overcome layers of convective inhibition), and those linked to its response (e.g. Mapes (2000) and Fletcher and Bretherton (2010). This can lead to numerical intermittency of CAPE-closed schemes across the timestep, as the CAPE-closure can create 40 36 enough convective inhibition to prevent triggering on the next timestep (e.g. Klingaman et al. (2017)). Furthermore closures based on CAPE desensitise convective activity to local variations in the vertical thermodynamic profile, mak-ing it harder for the convection scheme to respond in tandem with the resolved dynamics (Whitall et al. (2022)). In addition, the use of predefined profiles for entrainment leads to somewhat inflexible representations of the mass-flux 45 40 shape, making smooth transitions between different convective states hard to achieve.

These limitations, along with embedded decisions in the convection-scheme code about the levels of convective
 initiation, and the number of updraughts allowed at any one level, have motivated the development of an entirely new

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 convection scheme called CoMorph, Whitall et al. (2022). This scheme provides a much more flexible architecture that enables complexity to be built in as required, and with the aim of creating a scheme that is better connected to the underpinning physical processes that drive convective activity. 

CoMorph-A represents a first implementation of this scheme, designed for a global package, with the priorities of this version being to improve the coupling between convection and the resolved dynamics, as well as the consistency of interaction with other physics parametrisations. Numerous tests of the scheme under controlled forcing conditions have been conducted, for example in a Single Column Model (Whitall et al. (2022)), and for a range of case studies in the Idealised MetUM (a 3D, Cartesian, bi-periodic version of the Met Office Unified Model, Lavender et al. (2024)), with promising behaviour across a range of regimes. The coupling of the scheme to the dynamics has also been investigated (Daleu et al. (2023)), again highlighting the significant advances and potential of the new scheme. Zhu et al. (2023b) have also investigated the impacts of CoMorph-A in global simulations, focusing on the Indo-Pacific and Australian regions, finding significant benefits. 

This paper introduces the CoMorph-A and accompanying control package configurations of the Met Office Uni-fied Model (UM) in sections 2 and 3. To evaluate the impacts and performance of the new package, we follow the standard testing procedure as laid out in Walters et al. (2019), and evaluate the performance of the same configura-20 57 tion in both climate and numerical weather prediction (NWP). The climate evaluation in section 4 uses 20 years of simulation using AMIP specified sea-surface temperatures and sea ice. For NWP, while we have also run forecast only simulations, as in Walters et al. (2019), in section 5 we present verification from two 3-month long cycling forecast and analysis trials of the CoMorph-A package, so including 4DVar data assimilation (albeit without updating the error covariance and other assimilation statistics). 

#### CONTROL MODEL

The control configuration of the Met Office Unified Model (MetUM) used in this study is GA8GL9, which is based on GAL7.1GA7GL7 (Walters et al. (2019)) but incorporates newthe developments that were already intended for the 35 66 next release and that will be documented in detail elsewhere. That new trunk configuration, GAL8, GA8GL9 is the basis for the current operational NWP at the Met Office at the time of writing. The main changes since GAL7:1GA7GL7 are: 

- a package of changes related to model drag, described in Williams et al. (2020)
- revised parametrizations in the grid-scale microphysics for riming and the depositional growth of ice (Furtado and Field (2017))
- a new "multigrid" dynamical solver
- reduced drag at high wind speeds over the sea which improves forecasts of tropical cyclones at higher model resolutions than discussed here
- significant modifications to the existing massflux convection scheme

To give a brief description of GAL7.1GA7GL7, the ENDGame dynamical core uses a semi-implicit semi-Lagrangian for-mulation to solve the non-hydrostatic, fully compressible deep-atmosphere equations of motion (Wood et al. (1999)). The grid-scale microphysics scheme (i.e., that which operates on the gridded prognostic variables) is single-moment with prognostic rain and ice, based on Wilson and Ballard (1999). The parametrisation of the fraction of the grid box which is covered by cloud, and the amount and phase of condensed cloud water it contains, is the prognostic cloud fraction and prognostic condensate (PC2) scheme (Wilson et al. (2008a), Wilson et al. (2008b)). The parametrization of radiative transfer (solar, shortwave (SW), and thermal, longwave (LW)) uses the SOCRATES scheme (Edwards and Slingo (1996); Manners et al. (2015) The turbulence scheme uses a first-order closure from Lock et al. (2000), mixing adiabatically conserved heat and moisture variables, momentum and tracers, using non-local profiles of diffusion co-efficients within convective boundary layers and an explicit parametrization of entrainment across their top. In the control simulations the parametrization of sub-grid-scale transport of heat, moisture and momentum associated with cumulus clouds is a mass-flux convection scheme based on Gregory and Rowntree (1990). Three separate classes of convection are represented. First a diagnosis is made to determine whether convection is possible from the boundary layer, and its potential depth, using an adiabatic undilute parcel ascent, which then triggers either a shallow or deep convection scheme. Any remaining instability to moist ascent identified above the boundary layer and the top of any shallow or deep convection is handled by a mid-level scheme. Each component involves different parametrization of processes such as entrainment and detrainment from the plume, together with separate closures (based on the surface buoyancy flux for shallow convection and CAPE for mid-level and deep). 

Of the modifications to the GAL7.1GA7GL7 convection scheme used in our control, two are of most relevance here. The first, described in Willett and Whitall (2017), is to relate the entrainment rate in the deep component of 27 96 the scheme to the amount of recent convective activity, as measured by an advected prognostic, similar to Mapes 28 97 and Neale (2011). The source term for this is scaled on the surface convective precipitation rate, expanded to three dimensions using the current convective temperature increment profile, and decays with a 3 hour timescale. In this way the scheme is given a short term memory of previous convective activity, thereby allowing evolution of convective 32 100 clouds over finite time scales. The second significant change from GAL7.1GA7GL7 is to time-smooth the heat and 33 101 moisture increments from the convection scheme over a timescale of 45 minutes. This is to reduce the impact of strong time-step-level intermittency in the scheme (see section 4) on the resolved scale dynamics. Note that this increment smoothing is not used with the CoMorph-A package. 

Finally, note that most of these parametrizations have undergone significant modification between their original
 references, as above, and GAL7.1GA7GL7 and the reader is referred to Walters et al. (2019) for more detail.

# 106 3 | COMORPH-A PACKAGE

An important aspect of the new convection parametrization, CoMorph, is that it is directly coupled to, and hence strongly dependent on, other aspects of the model physics. In particular, the presence or otherwise of cloud strongly affects the initial sources of mass, and the turbulence statistics from the boundary layer scheme determine the initial characteristics of the plume. Initial development in a single-column version of the MetUM highlighted unwanted

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sources of intermittency and sporadic unrealistic behaviour at the time step and grid level. For the proper functioning
 of the CoMorph convection scheme, as well as for the overall performance, it was therefore necessary to make significant changes to several other aspects of the physical parametrizations that will be described in this section. Together
 these then form the CoMorph-A package.

### 115 3.1 | CoMorph convection scheme

11 <sup>116</sup> The control convection scheme is replaced by the entirely new scheme, CoMorph. Details are given in Whitall et al. 12 <sup>117</sup> (2022) but we give a summary here.

13 118 As in the control, the CoMorph scheme uses the massflux approach to parametrize the effects of subgrid cumulus 14 convection but CoMorph has an entirely new code structure, to allow greater flexibility, and many differences in 15 119 approach with additional physical processes included. There is no external closure in CoMorph. Instead, the massflux 16 120 17 121 initiated from any locally unstable model grid-level is proportional to the relative size of the local environment lapse 18 122 rate to that of a test parcel. Where the prognostic cloud fraction parametrization identifies a grid box as having partial cloud cover, separate initiation calculations are made in the saturated and unsaturated air and these are combined 123 20 using area weighting by the cloud fraction. 21 124

22 125 The rate of entrainment scales with an inverse parcel radius, i.e.,

$$\epsilon = \frac{\alpha_{ent}}{R} \tag{1}$$

where  $\alpha_{ent}$  is a dimensionless parameter currently set to 0.2. The parcel radius, R, is based on a turbulence length-28 126 scale in the parcel's source-layer given by  $R = a_L K_M / \sigma_w$ , where  $K_M$  is the momentum diffusivity in the boundary 29 127 30 128 layer scheme and  $a_L$  a constant taken to be 0.45;  $\sigma_w = (K_M/\tau)^{0.5}$  is a turbulent velocity scale where  $\tau$  is a turbulent 31 timescale, see Van Weverberg et al. (2016). Initial testing in the global model, however, found this simple formulation 129 32 was insufficient to remove instability fast enough under all circumstances and so an additional scaling was introduced, 33 130 increasing  $a_L$  based on an ad-hoc linear function of the previous time step's precipitation rate (up to a factor of 34 131 35 132 2 for a precipitation rate above approximately 35 mm per day). Qualitatively this can be viewed as allowing the 36 updraft size to respond to the organisation of convection via precipitation leading to the formation of cold pools and 133 37 subsequent initiation of larger-scale updraughts (Mapes and Neale (2011)). The parcel radius increases with height 38 <sup>134</sup> through entrainment by assuming the mass-flux is carried by a fixed number-flux of spherical thermals whose radii 39 135 40 136 increase consistent with the increase in volume of each thermal.

The detrainment rate is modelled as removing the non-buoyant part of an assumed distribution of buoyancy
within the bulk plume, informed by separate parcel ascent calculations for parcel mean and less dilute core properties.
Importantly, this detrainment calculation is solved implicitly, in which the convective heating of the environment (as
well as the other model increments through the timestep) is accounted for in the calculation of the parcels' buoyancy
and therefore the proportion of the buoyancy distribution that is detrained.

48 <sup>142</sup> The initiating parcel mean properties are calculated separately within each clear or cloudy sub-grid region, as with

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2 143 massflux initiation, and then averaged together weighted by their respective initiating massfluxes. The in-region prop-3 erties are taken as just saturated and neutrally buoyant and then given a perturbation,  $\phi'$ , linked to the parametrized 144 4 turbulent flux ( $\phi' = \alpha_t \overline{w' \phi'} / \sigma_w$  with  $\alpha_t = 1/3$  and  $\phi$  being temperature, water vapour mixing ratio and the horizon-145 5 tal wind components). Because the parcel core is representing the most extreme tail of the distribution of sub-grid 146 6 7 fluctuations, the core property perturbations are scaled up by a factor of 3. 147

As with the control, CoMorph uses a massflux representation of momentum transport by convection that also 148 includes the effects of cloud pressure-gradients on the flow within the cloud (e.g., Kershaw and Cregory (1997)). 149 10

11 150 Importantly, CoMorph also includes a novel in-parcel bulk microphysics scheme for the generation of hydrom-12 151 eteors. These hydrometeors are then passed on the model-level where they leave the parcel (through detrainment 13 and a fall-out flux) to the grid-scale microphysical variables. This can be achieved because, asAs in the control, the 152 14 model's grid-scale (Wilson-Ballard) microphysics scheme prognoses the mass concentration of each hydrometeor 15 153 16 154 species and so convection can now provides an additional source term for these prognostics. Their fall to the sur-17 155 face, and any interactions on the way down, is modelled by the Wilson-Ballard scheme which therefore improves 18 the consistency of treatment of precipitation between that generated through convective processes and that through 156 19 larger, resolved scale ones. Note that diagnostically this means there is no longer any separation into convective or 20 157 21 158 large-scale precipitation, just precipitation. In order for convection to modify rain mass and area fraction consistently 22 159 a prognostic precipitation area fraction is introduced and passed to the large-scale microphysics. In addition to cloud 23 160 liquid water, ice and rain, CoMorph itself includes graupel, which is not represented explicitly in the control model, 24 25 161 being largely a feature of deep convection which is entirely parametrized there, but passing CoMorph's graupel to the 26 162 large-scale's single ice variable led to far too excessive build-up of cloud ice in the tropics and so the large-scale micro-27 163 physics' prognostic graupel was also included in the CoMorph-A package (following the configuration already used in 28 the regional configurations of the UM, see Bush et al. (2020)). This initial implementation of a microphysics scheme 164 29 30 165 in CoMorph is relatively simple, prescribing a representative number concentration for each hydrometeor species in 31 166 order to calculate an effective particle size and thence fall speeds, accretion rates and other exchanges between hy-32 167 drometeor classes. Crucially this allows the parcel (and detrained air) to become supersaturated with respect to ice, 33 whereas the convection scheme in the control adjusts the parcel fully to ice-saturation when below the melting point 168 34 35 169 temperature.

36 170 CoMorph is also written in terms of mixing ratios of water vapour and condensate (while the control convection 37 <sub>171</sub> scheme uses specific quantities). This has then allowed all the model's physics schemes to move across to using mixing 38 ratios, consistent with the dynamical core, and so avoid conversions in moving between different parts of the model 39 <sup>172</sup> which allows global water conservation errors to be reduced to negligible levels. 40 173

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- 44 175 it is a single unified scheme for all convective regimes
- 45 176 it is closely integrated with the rest of the model physics, giving greater physical consistency

In summary, the key aspects of the CoMorph convection scheme are:

46 massflux is initiated through local moist instability leading to a much tighter coupling with the resolved dynamics 177 47 of the model 48 178

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2 179	• close attention has been paid to the numerical methods used, with substantial aspects solved implicitly, and this
3 180	leads to a smooth temporal behaviour
4 1 <sup>81</sup>	• it includes a microphysical representation of cloud process in its updraughts and precipitation is passed directly
Э 6 <sup>182</sup>	to the prognostic grid-scale microphysical variables instead of being rained out diagnostically, thereby improving
7 183	consistency of treatment of microphysical processes in the model
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10 184	3.2   Other model changes
11	2.2.1 Claud schome
12 185 13	S.2.1   Cloud scheme
14 186	The impact of triggering CoMorph from either saturated or unsaturated air makes it sensitive to the diagnosis of the
15 <sub>187</sub>	fraction of the grid box that is cloudy. In the process of developing CoMorph, several aspects of the PC2 cloud scheme
16	were discovered to be excessively sensitive to small perturbations. Largely motivated, then, by a desire to reduce the
17	timestep level variability of convection, the following changes were made to the PC2 cloud scheme:
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20	a to appure the cloud is consistent with the thermodynamic profiles at the point in the timester where these are
21	• to ensure the cloud is consistent with the diermodynamic promes at the point in the timestep where these are
22 191	passed to Comorph, the PC2 homogeneous forcing calculation from advection is moved from the end of the
23 <sup>192</sup>	timestep to just after the advection calculation (which is called before convection) and new calls to the PC2
24 193 25	initiation and checking algorithms are added before the call to CoMorph
25 <sub>194</sub> 26	• the "BiModal" scheme of Weverberg et al. (2021) is used for initiation of cloud within PC2 (because of its more
20 27	realistic treatment of subgrid variability), together with using that initiated cloud as a minimum cloudiness (cloud
28 <sup>196</sup>	fraction and liquid water content) at every grid-point on every timestep (in the control the cloud has to dissipate
29 197	completely before initiation is triggered which can lead to sudden large changes in cloud fraction)
30 198	• for consistency, the pressure change experienced by the environment as it undergoes convectively forced subsi-
31 22	dence is used to drive the homogeneous forcing of cloud, following the same method as for resolved advection
32 33 <sup>200</sup>	(Wilson et al. (2008a)), a process not treated explicitly in the control
34 201	• an implicit numerical method is introduced to calculate the PC2 term for the erosion of cloud, which depends on
35 <sub>202</sub>	the cloud fraction (see Morcrette (2012)), such that the erosion rate applied is consistent with the cloud-fraction
36 203	that will occur after the erosion increment has been added on
3/ 20 <sup>204</sup>	• the method of Morcrette (2020) is used to ensure that the PC2 cloud scheme produces reasonable magnitude
30 39 205	cloud fraction increments in the limit of small condensate amounts
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42 206	3.2.2   Cloud microphysics
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44 207 45	As discussed in section 3.1, the main changes to microphysics are to include graupel as a prognostic variable, which
46	allows for the explicit representation of a second, more dense ice category, and to include a fractional area of precipita-
47 209	tion so as to discriminate convectively generated rain that occupies a small fraction of the grid box from that generated
48 210	by large-scale processes. Another, more minor, improvement is to relax the shallow atmosphere assumption in the
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211 calculation of flux divergences, and so be consistent with the rest of the model, which improves local conservation of 212 water.

#### 3.2.3 Fountain Buster 213

In the absence of sufficient lateral subgrid mixing, the MetUM has been found susceptible to the formation of near-214 10 215 grid-scale flow structures that, combined with the lack of inherent conservation with semi-implicit semi-Lagrangian advection, leads to significant conservation errors. This problem can be illustrated by considering an idealised up-11 216 12 217 draught in a single grid-column with continuity requiring a convergent in-flow in lower levels. The interpolation of 13 218 the Arakawa C staggered horizontal wind fields onto scalar points, however, strongly reduces resolved lateral conver-14 gence from surrounding points into grid-scale vertically ascending plumes. As a result, the vertical transport of, for 15 219 example, water vapour away from the near surface is not compensated for by horizontal convergence of drier air into 16 220 17 221 the plume from the sides, with the result that moisture can be transported vertically ad infinitum. Typically this is then 18 222 manifested as extreme local precipitation accumulations. The use of a posteriori global conservation correction has 19 been found to reduce these errors significantly (Bush et al. (2020)), but substantial local errors have been found to 20 223 21 224 remain.

22 225 Here a post-hoc local correction is applied after the semi-Lagrangian scheme's interpolation to the departure 23 226 point, in the form of the linear up-wind advection increments that arise from the convergent part of the flow that 24 is removed when the horizontal wind components are interpolated to the scalar grid points for the departure point 25 227 calculation. In more detail, first, at each lateral cell face a measure of the (assumed missing) convergent stagnation 26 228 27 229 in-flow, S, is calculated. For example, for the west cell face of scalar point (i, j, k):

$$S_{\text{west}} = \frac{u_{i-1,j,k} - s_d u^{\theta}}{u_{i-1,j,k}}$$
 (2)

where  $u^{\theta}$  is a linear estimate of u at the scalar point. To ensure only the convergent part of the flow is accounted for, 33 230 34 231 we ensure  $0 < S_{west} < 1$ , and to avoid making increments in reasonably well-resolved flows the tuning factor,  $s_d$ , is 35 232 set to 2. 36

The fountain buster tendency for a scalar variable  $\chi$  is then given by first-order upwind advection scaled by S, 37 233 38 234 i.e.:

- $\begin{array}{lll} \frac{\Delta\chi_{i,j,k}}{\Delta t} & = & S_{\mathrm{east}}u_{i,j,k}^{f}\frac{\chi_{i+1,j,k}-\chi_{i,j,k}}{\Delta x} S_{\mathrm{west}}u_{i-1,j,k}^{f}\frac{\chi_{i,j,k}-\chi_{i-1,j,k}}{\Delta x} \\ & + & S_{\mathrm{south}}v_{i,j-1,k}^{f}\frac{\chi_{i,j,k}-\chi_{i,j-1,k}}{\Delta y} S_{\mathrm{north}}v_{i,j,k}^{f}\frac{\chi_{i,j+1,k}-\chi_{i,j,k}}{\Delta y} \end{array}$ (3)
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45 235 where the superscript f denotes the wind interpolated (vertically) to the face centre of the scalar point. In (3) a regular 46 cartesian grid has been used for simplicity but in the MetUM the appropriate metric variables are used. The fountain 236 47 buster is applied to all advected scalar variables, so all moisture variables, tracers and the thermodynamic variable (dry 48 237

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238 virtual potential temperature).

#### 3.2.4 **Turbulent mixing** 239

A deliberate choice in the Lock et al. (2000) boundary layer scheme was to interface with the convection scheme 240 at the lifting condensation level (LCL, diagnosed from the top of the surface layer), meaning that the boundary layer 241 10 242 scheme would be entirely responsible for surface-driven mixing at least up to the LCL, from where the convection 11 243 scheme would then be initiated and take over the transport into the free troposphere. With CoMorph, initiation can 12 244 now occur at any level and so we take the opportunity to relax the requirement that the boundary layer scheme mix 13 up to this definition of the LCL. We still use the Lock et al. (2000) undilute moist adiabatic parcel ascent method to 245 14 diagnose the potential for cumulus convection but, in those regimes, now diagnose the top of the surface-driven 15 246 16 247 turbulence diffusion profile as the point where the integral of diagnosed negative buoyancy flux is a fraction (0.05, 17 248 see Walters et al. (2019)) of the vertical integral of the positive flux. Because we assume any transport within clouds 18 249 will be carried by CoMorph, we use the unsaturated buoyancy flux in this calculation. 19

A second alteration to the boundary layer scheme is in the calculation of the buoyancy gradient used in the 20 250 21 251 Richardson number (Ri). This is revised to reduce the mixing from the Ri-dependent part of the scheme across 22 252 statically stable inversions, in particular where these cap a cloudy boundary layer. With the UM's Charney-Philips 23 24 <sup>253</sup> vertical grid staggering, Ri is required on the same grid-level as scalars (referred to as  $\theta$ -levels), where the vertical gradient of horizontal momentum naturally falls. In the control, the gradients of liquid-ice static energy temperature 25 254 26 255 and total water content are interpolated to  $\theta$ -levels and that  $\theta$ -level's cloud fraction used to obtain the buoyancy 27 <sub>256</sub> gradient. It is found that at cloudy  $\theta$ -levels below a strong inversion this interpolation of the gradients, combined 28 29 <sup>257</sup> with a saturated buoyancy calculation, can lead to apparent instability. In the CoMorph-A package we calculate the 30 258 buoyancy gradient locally (on  $\rho$ -levels) and interpolate that to the  $\theta$ -levels. To calculate the saturated contribution to 31 259 the buoyancy gradient on  $\rho$ -levels, an estimate of the vertical fraction of the grid containing saturated air is made from 32 260 interpolating the supersaturation in the adjacent  $\theta$ -levels. In this way the contribution from the  $\rho$ -level with strong 33 gradients is typically largely unsaturated, leading to strong stability, which then remains stable when averaged with 34 <sup>261</sup> the saturated  $\rho$ -level with weak gradients below. 35 262

36 263 As with the Fountain Buster, the more active dynamics in the tropics arising from more organised convection with CoMorph also motivates enabling the Leonard terms (Hanley et al. (2019)), which include extra subgrid vertical fluxes 264 that account for the tilting of horizontal flux into the vertical by horizontal gradients in vertical velocity. Including these 39 <sup>265</sup> terms helps to weaken a few occurrences of excessively strong resolved updraughts and precipitation in organised 40 266 41 267 convective structures.

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#### Tuning of the top-of-atmosphere radiation 3.3 268

46 As with any major model developments, the initial simulations showed significantly worse biases than the control, 269 47 both locally and globally in the top-of-atmosphere radiation. To reduce excessively weak outgoing longwave radiation 48 270

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2 271 (OLR) in the tropics, we increased the fall speeds for cloud ice by 20% (which increases the global mean OLR by around 3 1.5 Wm<sup>-1</sup> and reduces the spatial root-mean-square error (rmse) against satellite climatology (see section 4) by 0.3 272 4 Wm<sup>-1</sup>) and also removed a parametrization in PC2 that increased the ice cloud fraction through vertical wind shear 273 5 that was not especially well justified physically (which increased the global mean OLR, by 0.5 Wm<sup>-1</sup>, and reduced 274 6 the rmse by  $0.3 \text{ Wm}^{-1}$ ). In the shortwave (SW), the cloud forcing was initially too strong across most of the tropics. 7 275 8 Several options are available that can reduce the reflectivity of clouds and we have made two revisions. The first was 276 9 to the parameters in the Liu et al. (2008) spectral dispersion of the cloud droplet size distribution. As described in 277 10 Mulcahy et al. (2018), the cloud droplet spectral dispersion as implemented in the control is given by 278 11

$$\beta = a \left(\frac{L}{N_d}\right)^b \tag{4}$$

16 279 where L is the cloud liquid water content and  $N_d$  the cloud droplet number concentration. In the control a = 0.07 and 17 18 280 b = -0.14 while in the CoMorph-A package we use a = 0.093 and b = -0.13. These changes, that are small compared to the spread in the Liu et al. (2008) data, resulted in a reduction in global mean reflected SW of around 1.5 Wm<sup>-1</sup>, 19 281 20 282 whilst having almost no impact on OLR. Regionally the SW impact was strongest broadly across the marine sub-tropics 21 283 and gave a reduction in spatial rmse of around 1 Wm<sup>-1</sup>. Our second tuning, to further reduce the reflectivity of clouds, 22 was to increase a parameter in the fractional standard deviation of liquid clouds (equal to the standard deviation of 284 23 cloud water content in a grid box divided by its mean value, see Hill et al. (2015)) from 1.6 to 1.65, which reduced the 24 285 25 286 global mean reflected SW by around 0.7 Wm<sup>-1</sup> and the spatial rmse by 0.25 Wm<sup>-1</sup>. Note, though, that the impacts 26 287 of these last two changes would certainly be different if implemented in the reverse order. 27

It is important to remember that this tuning is a critical step in any new physics package and was, of course, also 28 288 done for the control configuration. Hence the impacts discussed in section 4 should be viewed as resulting from the 29 289 30 290 CoMorph-A package as a whole.

#### 3.4 Cost 291

There is a marginal increase in the CPU time with CoMorph-A of roughly 5% over the control of which a substantial 292 part will be the new prognostic graupel - tests with just the convection scheme reverted to that in the control show 37 293 38 294 a similar increase indicating the CoMorph convection scheme itself is cost neutral.

#### **RESULTS IN AMIP** 4 295

44 296 We evaluate the climatology of the CoMorph-A package using AMIP simulations at N96 resolution (around 135 km 45 297 grid spacing in mid-latitudes). Given the main change is to the convection parametrization, we start by looking in Fig. 1 46 at the impact on the annual mean precipitation. The most systematic change is to have increased rain over tropical 298 47 land, including the maritime continent, most of which is beneficial compared to the Global Precipitation Climatology 48 299

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Project (GPCP, Adler et al. (2003)). Particularly relevant for the Maritime Continent, a significant issue noted by operational meteorologists with the control simulation in the tropics is precipitation in convective weather systems abruptly stopping at the coastline as they move from sea to land. This issue is much improved in CoMorph-A. Especially over the Pacific Ocean there is a sharpening of the ITCZ, giving enhanced rainfall on the equator with reductions to the north and south, with mixed impact.

8 More detailed analysis shows the characteristics of the model's tropical precipitation at local time and space scales 305 9 are altered beyond recognition. Fig. 2 quantifies the probability of grid point precipitation rates at each timestep given 10 <sup>306</sup> the rate at the previous timestep. In the control model there is a strong signal along the axes, indicating a high proba-11 307 12 308 bility of no precipitation on one timestep if there was precipitation on the previous, and vice versa - this illustrates the 13 309 long-standing, Klingaman et al. (2017), time-step level intermittency in the control. With CoMorph-A this timestep in-14 termittency is completely removed and shows a far more plausible temporal coherence with high probabilities around 15 310 the one-to-one line. Fig. 2 shows that the local characteristics of tropical precipitation are altered beyond recognition, 16 311 17 312 with the long-standing Klingaman et al. (2017) time-step level intermittency in the control completely removed in 18 313 CoMorph-A, which shows a far more plausible correlation. This continuity between time steps is a direct result of the 19 use of improved numerical methods, as discussed in section 3, and the implicit solution for detrainment in CoMorph, 20 314 in particular, as this will tend to maintain similar convective inhibition into the next time step. Interestingly, after 21 315 22 316 averaging over 3 hours and 2x2 grid points for comparison with the GPM observations (Global Precipitation Measure-23 317 ment (GPM) Multi-satellitE Retrievals (IMERG) precipitation data V06B, Tan et al. (2019)), the models' precipitation 24 25 <sup>318</sup> characteristics are far more similar. This convergence is similar to that seen for a range of MetUM configurations at various resolutions in Martin et al. (2017) and similar to the results for a number of CMIP5 models (except for those 26 319 27 320 with particularly coarse resolution) in Klingaman et al. (2017). There is, nevertheless, a suggestion of more temporally 28 321 coherent heavy rain with CoMorph-A, closer to GPM, albeit still insufficient. It is also interesting to reflect that many 29 30 <sup>322</sup> of the geographical biases in tropical precipitation seen in Fig. 1 are remarkably similar, despite entirely replacing the convection scheme, although many of these simply reflect the regions with the heaviest rainfall. 31 323

The strength of propagating tropical waves is strongly improved, see Fig. 3, both for Kelvin waves (as marked on the figure) and the Madden-Julian Oscillation (MJO, the region of strong observed eastward-propagating 30–60 day variability at wave numbers 1 to 3). We believe this tighter coupling between CoMorph and the resolved dynamics arises because the initiation of massflux can respond to locally generated moist instability, but more detailed investigation will be the subject of future work.

Fig 4 shows a large increase in the integrated tropical water vapour that all but removes a dry bias in the control,
largely occurring in the upper troposphere. Around a half of that moistening comes from the Fountain Buster scheme,
while the rest is likely due to passing the precipitation into the "large-scale" microphysics that then represents its
evaporation as it falls to the surface more realistically (although to confirm this would require adding some form of
direct precipitation to the surface within CoMorph, which is beyond the scope of this work). This moister troposphere
may also contribute to the improvement in the strength of tropical waves, see Fig. 3 (Zhu et al. (2023a)).

46<br/>47335There are also substantial changes to the clouds in the model, as seen in the top-of-atmosphere cloud radiative48<br/>48<br/>336effect (CRE, which is the clear-sky minus the total radiative flux). The longwave CRE, Fig. 5, is much stronger across the

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2 337 tropics and especially over tropical land, which all but eliminates a serious underestimate in the control. This is due to 3 a large increase in the tropical ice cloud amount (both mass and cloud fraction) as can also be seen in the comparison 338 4 with CALIPSO observations over central Africa in Fig. 6, likely arising from the handling of precipitation sedimentation 339 5 by the grid-scale microphysics. However, like the control, there is less cloud than observed by CALIPSO between 5km 340 6 7 and 8km altitude suggesting that the congestus phase remains poorly simulated. The increase in optically thick cirrus 341 8 342 means shortwave CRE (Fig. 7) is also enhanced somewhat over tropical land, but the biggest impacts here are in the 9 subtropics and especially over the eastern Pacific and Atlantic. Climatalogically, these are areas dominated by stratocu-343 10 mulus clouds. The geographical pattern of the change from including the CoMorph-A package matches remarkably 344 11 well with the error in the control, although the end result is for the clouds to be somewhat too reflective over most of 12 345 13 346 the tropical oceans. Part of the increase in stratocumulus arises from the implementation of BiModal initiation in the 14 PC2 cloud scheme and the revised calculation of Ri, but around half comes from CoMorph itself, especially further 347 15 west from the coastline where shallow cumulus clouds form under the lifting stratocumulus. Detailed single column 16 <sup>348</sup> model analysis of this transition region (not shown) indicates that the combination of the smoother temporal evolution 17 349 18 350 of CoMorph (as in Fig 2) and more sensitive (implicit) detrainment allow a smoother evolution of the stratiform cloud 19 layer as it rises away from the coast. Consistent with the improved top-of-atmosphere radiative fluxes, we also find 351 20 21 352 the net surface energy fluxes (that are important for coupled ocean modelling) are significantly improved, and across the tropics and subtropics especially, compared to the observed estimate from Liu et al. (2015) (not shown). 22 353

23 354 A form of convective "memory" was implemented in the control model (in GA8GL9 relative to GA7GL7, as de-24 355 scribed in section 2) to delay the development of deep convection and help address the poor representation of the 25 26 356 diurnal cycle of precipitation over land. However, consistent with other studies (Willett and Whitall (2017); Tao and 27 357 et al (2024)), we find GA8GL9 still has issues in simulating the correct timing of the diurnal cycle over many areas of 28 358 the globe (see Fig. 8). CoMorph-A does not have such a memory function and CoMorph-A does not contain a form 29 359 of "memory", unlike the control. Consequently, it has a tendency to have the diurnal peak in rainfall intensity over 30 tropical land around local noon or early afternoon, compared to the observed late evening peak (Fig. 8). Over some 31 360 32 361 regions (e.g. Africa) this is earlier and hence detrimental compared with the control, whilst over others (e.g. S. E. Asia, 33 <sub>362</sub> as also seen in Zhu et al. (2023b)) CoMorph-A is an improvement on the control, although still considerably earlier 34 363 than observed. Diurnal cycle experiments using the idealised MetUM (Lavender et al. (2024)) also show too early 35 initiation and development to deep convection using CoMorph-A compared to high resolution simulations, although 36 364 improved relative to the control. 37 365

### 40 366 5 | **RESULTS IN NWP**

<sup>43 307</sup> The NWP trials (cycling forecasts and data assimilation) are performed at somewhat higher, N320 (~ 40 km in mid<sup>44 308</sup> latitudes), resolution than the climate simulations presented in section 4. A wide variety of metrics are monitored and
<sup>45 309</sup> these are summarised in the "scorecards" shown in Fig. 9. These illustrate the change in root-mean-square error (rmse)
<sup>46 370</sup> against each trial's own analysis (scorecards are also produced against observations and independent global analyses
<sup>48 371</sup> from ECMWF but those look very similar so are not shown). The majority of metrics show reductions in rmse with

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2 372 CoMorph-A, with somewhat better overall performance in the boreal summer than winter (overall reductions of 1.9% 3 and 0.5% respectively). Some degradation is seen in a number of fields in the first two days that may arise from not 373 4 updating the error covariance statistics in the trials. One field that shows degradation throughout the forecasts in both 374 5 summer and winter, though, is the tropical temperature at 850 hPa. Fig. 10 shows the geographical distribution of the 375 6 7 change in rmse at day 6, which has large positive values across most of the tropical oceans but especially so, for these 376 8 377 austral summer trials, in the sub-tropical south-eastern Pacific and Atlantic close to the edges of where CoMorph-A 9 gives more stratocumulus cloud layers (the NWP trials show a similar increase to that shown to be beneficial in the 378 10 AMIP simulations in Fig. 7). Fig. 10 also shows that the standard deviation of temperature at 850 hPa in the analyses is 379 11 increased in CoMorph-A, and the day 6 forecasts in CoMorph-A have similarly more variable temperatures at this level 12 380 13 381 (not shown). The more persistent stratocumulus clouds with CoMorph-A will result in a sharper temperature inversion 14 382 at the top of the boundary layer. Furthermore, on the western edge (and further downwind) of the stratocumulus, 15 there is more variability of the cloud cover as it transitions to shallow cumulus. These transitions result in variations in 383 16 the height of the inversion which will be manifested as increased temperature variability and rmse at this level, even 17 384 18 385 if the average inversion height or boundary layer temperature were accurately forecast (the 1000 hPa temperature 19 386 rmse is actually improved or neutral over tropical oceans, not shown). 20

A long standing systematic error is a slow bias in the winter extra-tropical jet. Fig. 11 shows an increase in the wind 21 387 22 388 speed from CoMorph-A, reducing this error. It is often the case that increasing the wind speed will also increase the 23 rmse due to a "double penalty" if a wind feature is geographically displaced. It is notable, therefore, that CoMorph-A 389 24 25 <sup>390</sup> reduces the slow bias with an overall neutral impact on rmse for upper level winds (Fig. 11 & Fig. 9).

26 391 Further analysis of the trials has included tracking the location of extra-tropical cyclones through the maximum in 850 hPa vorticity (Hodges (1995)). Whilst the error bars overlap and hence cannot be regarded as significantly 392 different, the errors in these tracks are reduced in CoMorph-A and experience suggests the fact the signal is consistent 29 <sup>393</sup> in the two seasons is notable (Fig. 12). 30 394

Finally, between 25 June to 31 October 2021, atmosphere-only forecasts with the CoMorph-A package were run 395 daily at the operational resolution of N1280 from Met Office operational analyses at 0 UTC. Verification of tropical 33 <sup>396</sup> 34 397 cyclones from these forecasts in Fig. 13 shows CoMorph-A gave substantial deepening and improvement to wind 35 398 speeds. Whilst the central pressure in these simulations is now lower than observed, they are atmosphere-only and we would expect slower-moving storms to be weaker when coupled to an ocean model in operational systems. 399

#### CONCLUSIONS 6

The results from a range of global simulations have been presented for a large package of revised parametrization 43 401 44 402 schemes, centred around the entirely new CoMorph massflux convection scheme. The key components of CoMorph 45 403 are that it has a physically-based flexible formulation that allows it to represent all convective regimes and is tightly 46 coupled to the rest of the model, both the other parametrization schemes and the dynamics. The main impacts of this 404 47 overall CoMorph-A package in standard global test simulations are found to be: 48 405

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406 407	• removal of the timestep level intermittency of convection that has been in all configurations of the MetUM since its inception
408	improved tropospheric humidity over tropical land
409	improved tropical variability, including Kelvin waves and MJO
410	surface fluxes improved in many regions, important when coupling to an ocean model
411 412	<ul> <li>increased high cloud over tropical land where it has always been lacking and hence marks a significant improve- ment</li> </ul>
413	• increased low cloud over tropical oceans which is generally better but optically too bright everywhere (although
414 415	tuned for the global mean net top-of-atmosphere radiative flux, in order to balance weak clear sky out-going longwave radiation)
416	<ul> <li>improved NWP performance, including reduced tropical and extra-tropical cvclone errors</li> </ul>
417	some degradation of the diurnal cycle of tropical precipitation
418	Given the extent of new developments within the CoMorph-A package, this represents impressive performance for
419	a first implementation of a new convection scheme. Future work will focus on improving the diurnal cycle, through
420	for example inclusion of a second updraught to represent, separately, convection forced from cold pools, and on
421	CoMorph's applicability to higher resolutions, where the larger scales of convection begin to be resolved.
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434	munity for many stimulating discussions.

# 45 <sup>₄₃₅</sup> Data availability statement

47 436 The data that support the findings of this study are available from the corresponding author and Met Office co-authors 48 437 upon reasonable request.

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FIGURE 2 Histograms of the probability of bins of precipitation, measured at the time step and grid point level (top
 row) and over 3 hourly and 2x2 grid box averages (bottom row), aggregated over all grid points in the equatorial Indian
 Ocean from N96 AMIP simulations. The dashed lines show the one-dimensional histogram of the binned precipitation,
 using the right-hand vertical axis.

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FIGURE 4 Annual mean total column water vapour from N96 AMIP simulations, and compared to ERA40 (Uppala et al. (2005)) in the bottom row.

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**FIGURE 5** Longwave cloud radiative effect compared to Clouds and the Earth's Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) dataset (Loeb et al. (2009)) from N96 AMIP simulations



FIGURE 6 Histograms of height vs 532nm lidar backscatter ratio over central Africa (15°E-30°E, 20°S-10°N),
 showing climatologies of CALIPSO observations (Winker et al. (2002)) and simulated climatologies of CALIPSO data
 from 20-year N96 atmosphere/land-only climate simulations from the control and CoMorph-A

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FIGURE 8 Local time of maximum in the diurnal harmonic of precipitation during JJA compared to TRMM 3B42 observations



FIGURE 9 Scorecards against own analysis from the summer (left) and winter (right) NWP trials. Green upward triangles indicate reductions in root-mean-square error (rmse) and purple downward triangles increases, with the size of each scaled by the magnitude of change. Parameters are evaluated in the Northern Hemisphere (NH, north of 18.75° N), the tropics (TR, within 18.75° of the equator), the Southern Hemisphere (SH, south of 18.75° S), Europe (Euro) and UK (UK4 and UKIndex). The parameters are pressure at mean sea level (PMSL), vector wind (W), tempera-ture (T) and geopotential height (Z) at various pressure levels given in hectopascals (such that NH\_T250, for example, measures the change in rmse temperatures at 250 hPa in the northern hemisphere region). The columns represent forecast ranges from 6 h (T+6) to 7 days (T+168). 



**FIGURE 10** Change in root-mean-square error against own analysis of day 6 temperature at 850 hPa (top) and change in standard deviation of the analyses (bottom), from the winter NWP trials



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The performance of the CoMorph-A convection package in global simulations with the Met Office Unified Model

A. P. Lock, M.Whitall, A. J. Stirling, K. D.Williams, S. L. Lavender, C. Morcrette, K. Matsubayashi, P. R. Field, G. Martin, M.Willett, J. Heming

The new CoMorph-A package of physical parametrizations in the Met Office Unified Model is documented and shown to perform well in global configurations. The main component is an entirely new convection scheme, CoMorph, but it also includes significant changes to the cloud, microphysics and boundary layer parametrizations. Biases in radiative flux climatologies are significantly reduced (top panel for the control annual mean longwave cloud radiative forcing, bottom for CoMorph-A), tropical and extratropical cyclone statistics are improved, the MJO and other tropical waves are strengthened and it also improves overall scores in NWP trials.

