The Effects of Latent Heat Release on the Climate of an Aquaplanet Model

RUTH GEEN

Space and Atmospheric Physics Group
Department of Physics
Imperial College London

Submitted in part fulfillment of the requirements for the degree of Doctor of Philosophy in Physics of Imperial College London and the Diploma of Imperial College

November 2015
I declare that the work presented in this thesis is my own, except where explicitly indicated otherwise.

The copyright of this thesis rests with the author and is made available under a Creative Commons Attribution Non-Commercial No Derivatives licence. Researchers are free to copy, distribute or transmit the thesis on the condition that they attribute it, that they do not use it for commercial purposes and that they do not alter, transform or build upon it. For any reuse or redistribution, researchers must make clear to others the licence terms of this work.

Ruth Geen, November 2015.
The Effects of Latent Heat Release on the Climate of an Aquaplanet Model

Abstract

As the atmosphere warms under climate change, it will hold more moisture. Latent heat released as water vapour condenses provides an important contribution to the atmospheric heat budget, affecting stability and providing complex feedbacks. Consequently, theories for the general circulation of the atmosphere proposed based on dry dynamics may not apply in moist simulations. In order to understand the possible changes to the Earth’s atmospheric circulation as the climate warms, a deeper understanding of these feedbacks is required.

Changes to the atmospheric thermal structure and circulation as humidity is increased have been explored in an intermediate complexity general circulation model. To provide a reference climate more comparable with that of previous studies, and of the real world, a simple parameterisation of shortwave and longwave radiative transfer has been developed, which compares favourably with existing simple radiation schemes. Experiments have then been performed with fixed optical depths in which the moisture content of the model is varied.

In the zonal mean, increasing moisture content results in an increase in static stability throughout the atmosphere. Consequent changes to the Hadley cell, zonal jets, and storm track have been analysed using simple theories, and by comparison with an experiment in which the sea surface temperature in the tropics is increased. This reveals that the majority of the effects of increased moisture content on the circulation are generated by low latitude warming.

The simulations further reveal stronger midlatitude poleward transport of moist static energy as saturation vapour pressure is increased, and an unexpected increase in sensible heat transport in the cold sector of storms. A mechanism for the latter is proposed related to the environmental static stability against which the system develops. The experiments also suggest changes to the rate of conversion of available potential energy to eddy kinetic energy as moisture content increases.
Acknowledgments

I would firstly like to thank my supervisors, Dr Arnaud Czaja and Prof Jo Haigh, for all of their help, advice, and encouragement throughout my PhD. I am very grateful for both the amount that I have learnt in the past three years, and for all of the espressos, chocolates, and kind words I have received in the process. I would also like to thank Dr Jean-Michel Campin and Dr David Ferreira for their help in setting up the model and in answering my questions about the code.

I have greatly enjoyed my time in SPAT, and would like to thank Team Arno, 708, the climbing crew, and really everyone who has brightened up my day with lunch and tea breaks, or surprise Italian conversation lessons, and made sure that work isn’t all about work. I particularly thank Josh Chadney and Amy Seales for their friendship and support over the last two years. I’ve also been lucky to have some wonderful housemates, both in Golders Green and Kentish Town, who I thank for all the lovely meals and fun times. Finally I thank my family for all of their support, and for their absolute confidence that I am clever enough to be a doctor.
## Contents

1 Introduction .................................................. 23
  1.1 Aquaplanets and the general circulation ................. 24
    1.1.1 Climate on an aquaplanet .......................... 25
    1.1.2 Equations determining the circulation ............ 27
  1.2 Radiative processes ....................................... 32
    1.2.1 Radiation and climate ............................. 32
    1.2.2 Radiative transfer .................................. 35
  1.3 Latent heat release and zonal mean climate ............. 37
    1.3.1 Static stability .................................... 37
    1.3.2 The Hadley Cell .................................... 41
    1.3.3 The jets and the zonal momentum budget .......... 46
  1.4 Latent heat release and baroclinic eddies ................ 48
    1.4.1 Growth and decay of storms ....................... 49
    1.4.2 Energetics .......................................... 52
    1.4.3 Eddy heat fluxes .................................. 55

2 Set-up and Experiments ...................................... 59
  2.1 The MITgcm .............................................. 60
    2.1.1 Dynamical core ..................................... 60
    2.1.2 Physics parameterisations .......................... 62
  2.2 Experiments .............................................. 66
  2.3 Climatology ................................................ 70
  2.4 Radiation modelling ...................................... 75
    2.4.1 SBDART ............................................. 75
    2.4.2 Simple radiation schemes ........................... 76

3 Parameterisation of the Radiative Effects of Water Vapour .. 79
  3.1 SBDART results and testing of longwave models ........ 80
  3.2 Shortwave parameterisation ............................... 84
  3.3 Longwave Scheme .......................................... 88
3.4 Incorporation in the MITgcm ........................................... 94
3.5 Conclusions .............................................................. 100

4 Effects of latent heat release on the zonal mean state .......................... 103
4.1 “Moisture content” experiments ........................................... 104
  4.1.1 Diabatic heating effects of moisture ..................................... 104
  4.1.2 Temperature and stability .................................................. 110
  4.1.3 The meridional overturning circulation .................................... 117
  4.1.4 The jet streams and momentum budget ...................................... 124
  4.1.5 Baroclinic instability and wave propagation ............................. 126
  4.1.6 Summary ........................................................................ 131
4.2 Comparison with increased tropical SSTs ......................................... 131
4.3 Eddy length scales .................................................................. 136
4.4 Conclusions ........................................................................ 142

5 Changes to eddy behaviour with increased humidity ............................. 147
5.1 Storm systems on the aquaplane .................................................. 148
  5.1.1 Storm climatology: Snapshots .................................................. 149
  5.1.2 Storm climatology: Composites ................................................. 160
5.2 Response of heat transport by extra-tropical waves ............................ 165
  5.2.1 Moist static energy transport .................................................... 165
  5.2.2 Sensible heat transport ........................................................... 166
  5.2.3 Mechanism ........................................................................ 168
5.3 The effect of latent heat on the energy cycle ..................................... 172
5.4 Conclusions ........................................................................ 173

6 Conclusions ........................................................................ 177
6.1 Summary of Results .................................................................. 178
  6.1.1 Radiative parametrisation ......................................................... 178
  6.1.2 Effect of latent heat release on the mean state ......................... 180
  6.1.3 Effect of latent heat release on midlatitude eddies ....................... 183
6.2 Future work .......................................................................... 185
  6.2.1 Further experiments with a slab ocean ....................................... 185
  6.2.2 Extension to the radiation scheme ............................................. 186
  6.2.3 The effect of latent heat on the energy cycle ............................. 187

References ........................................................................ 189
Commonly Used Symbols

$\Gamma$ Lapse rate ($Km^{-1}$)
$\mathcal{F}$ Friction ($ms^{-2}$)
$\mathcal{Q}$ Diabatic heating ($K/s$)
$\mu_s$ Saturation mass mixing ratio ($kg/kg$)
$\Omega$ Planetary angular rotation speed ($rad/s$)
$\omega$ Vertical wind speed ($Pa/s$)
$\Phi$ Geopotential ($m^2s^{-2}$)
$\phi$ Latitude
$\Psi$ Mass streamfunction ($kg/s$)
$\rho$ Density ($kgm^{-3}$)
$\sigma$ $p/p_s$
$\tau$ Optical depth
$\theta$ Potential temperature (K)
$a$ Planetary radius ($m$)
$c_p$ Specific heat capacity of dry air at standard temperature and pressure ($JK^{-1}kg^{-1}$)
$E$ Evaporation rate ($s^{-1}$)
$e_{s0}$ Reference saturation vapour pressure ($Pa$)
$e_s$ Saturation vapour pressure ($Pa$)
$F$ Eliassen-Palm flux
$f$ Coriolis parameter ($s^{-1}$)
$g$  Acceleration due to gravity ($ms^{-2}$)

$L$  Specific latent heat ($Jkg^{-1}$)

$N$  Brunt-Väisälä frequency ($s^{-1}$)

$P$  Precipitation rate ($s^{-1}$)

$p$  Pressure (Pa). $p_a$ indicates surface pressure, $p_0$ indicates reference surface pressure (1000 hPa).

$q$  Specific humidity ($kg/kg$)

$R_a$  Specific gas constant for dry air ($JK^{-1}kg^{-1}$)

$R_v$  Specific gas constant for water vapour ($JK^{-1}kg^{-1}$)

$SST$  Sea surface temperature

$u$  Zonal wind speed ($m/s$)

$v$  Meridional wind speed ($m/s$)
List of Figures

1.1 APE multi-model means. Top row: zonal average of zonal wind speed, m/s, temperature, K, and meridional wind speed, m/s. Bottom row: zonal average of vertical wind speed, mb/day, specific humidity, g/kg, and relative humidity. Adapted from Williamson and Coauthors [2012] (Figure 4.15). .................................................. 26

1.2 Zonal average of absorbed shortwave radiation (ASR) and outgoing longwave radiation (OLR) as a function of latitude, with the shading indicating the surplus in the tropics and deficit at high latitudes. Data is from the CERES experiment [Wielicki et al., 1996]. Adapted from Donohoe and Battisti [2012] (Figure 1). ©American Meteorological Society. Used with permission. .................................................. 33

1.3 Clear sky absorption spectra for common, strongly absorbing gases in the atmosphere. Taken from Figure 3.14, Chapter 3 of Andrews [2010]. ©Cambridge University Press 2000. .................................................. 34

1.4 Meridional temperature difference across the storm track (defined to be between 20° and 70°N at 500mb) vs bulk moist stability at 50°N for NCEP reanalysis data. Adapted from Juckes [2000] (Figure 6). ©American Meteorological Society. Used with permission. .................................................. 39

1.5 Vertical cross sections of (a) the zonal wind speed and (b) the mean meridional streamfunction, taken from annual-mean NCEP-NCAR reanalysis data between 1979 and 2001. Dashed contours indicate negative values, heavy contour shows the zero line for the zonal wind. (a) Contour interval is 5 m/s for positive and 2 m/s for negative values. (b) Contour interval is $2 \times 10^{10}$ kgs$^{-1}$, with only odd contours plotted. Taken from Dima et al. [2005] (Figure 1). ©American Meteorological Society. Used with permission. .................................................. 41
1.6 Vertical cross sections of the dominant terms in the angular momentum budget (Equation 1.17), taken from annual-mean NCEP-NCAR reanalysis data between 1979 and 2001. (a) $\overline{r}(f - \frac{\partial u}{\partial y})$. (b) $\frac{\partial v}{\partial y}$. Dashed contours indicate negative values, contour interval is $1 \times 10^{-5}$ ms$^{-2}$ Adapted from Dima et al. [2005] (Figure 6). ©American Meteorological Society. Used with permission.

1.7 Horizontal composites of the development of a midlatitude storm at (top to bottom) 300, 700, and 925 hPa. Data is from ERA-Interim reanalysis for DJF of 1989-2009. Left to right, columns show 48h and 24h before maximum intensity, and maximum intensity. Solid contours indicate geopotential height, m, dashed lines equivalent potential temperature, K, and arrows wind vectors relative to the system. Shading in the middle row indicates vertical velocity, Pa/s, and in the top row divergence. Taken from Dacre et al. [2012] (Figure 4). ©American Meteorological Society. Used with permission.

1.8 Schematic of the air masses in an extratropical storm, illustrating the motions of the warm and cold conveyor belts and the dry intrusion. Taken from Catto et al. [2010] (Figure 1). ©American Meteorological Society. Used with permission.

1.9 Eddy kinetic energy vs mean available potential energy for simulations with a moist model varying longwave optical thickness (circles) and insolation gradient (asterisks). Taken from O’Gorman and Schneider [2008b] (Figure 5). ©American Meteorological Society. Used with permission.

1.10 Composite pressure vs longitude colour map of meridional heat transport, W/1000hPa, for extreme $v^iH'$ events. Diagonal striping indicates equatorward heat transport. Cross hatching indicates regions that are not statistically significant at the 99% confidence level. Continuous contours show meridional velocity anomalies (spacing 5 m/s), dashed contours moist static energy anomalies (spacing 3 K). Thicker contours correspond to negative values. (a) and (b) show events corresponding to positive and negative meridional velocity anomalies at the location of the $v^iH'$ maximum respectively. The data cover latitudes north of 30°N, and includes DJFs from December 1989 to February 2011. Taken from Messori and Czaja [2015] (Figure 3). ©2015 Royal Meteorological Society. Used with permission.
2.1 (left) Potential temperature, K, and (right) specific humidity, g/kg, profiles used for model initialisation. .................. 67

2.2 Pressure and area averaged kinetic energy for the first 3 years of the wet experiment. Solid line indicates the time mean over the 10 years used for analysis. .................. 67

2.3 Sea surface temperatures used in experiments. Solid line shows control profile, dashed line shows perturbation used in tropheat experiment. .... 68

2.4 Zonal mean climatology of wet experiment. Top row: zonal wind speed, m/s, temperature, K, and meridional wind speed, m/s. Bottom row: vertical wind speed, hPa/day, specific humidity, g/kg, and relative humidity. Data is averaged over 10 years and over both hemispheres. ............... 71

2.5 Vertical integral of the terms in the angular momentum budget (Equation 1.40) for each run over all model levels. Solid line represents $C_{total}$, dot-dashed line $C_{zonal}$, dashed line $C_{eddy}$, cyan line (-1×) surface stress, and red line budget residual. Horizontal dashed line indicates the global average of the surface wind stress. .................. 72

3.1 (left) Temperature, K, and (right) specific humidity, g/kg, profiles used for comparison of radiation schemes. .......................... 81

3.2 (left) Longwave and (right) shortwave heating rates, K/day, calculated using SBDART for the reference profiles shown in Figure 3.1. Note the different contour interval between the panels. .......................... 82

3.3 Longwave heating rates, K/day, for the reference profiles shown in Figure 3.1 based on parameterisations from (left) Frierson et al. [2006] and (right) Byrne and O’Gorman [2013]. .......................... 83

3.4 Zonal mean zonal wind, m/s, (shading) and temperature, K, (contours) when the MITgcm is run with the parameterisations of (left) Frierson et al. [2006] and (right) Byrne and O’Gorman [2013]. .......................... 84

3.5 $d\tau/d\sigma$ as a function of $q$, g/kg. Evaluated from SBDART data using Equation 3.2. .......................... 86

3.6 Water vapour absorption coefficient, $b$, as a function of optical depth, $\tau$. Evaluated from SBDART data using Equation 3.3. Red line indicates the fit to a tropical profile used in the MITgcm (Equation 3.4). Blue line shows fit to a polar profile to be tested for comparison (Equation 3.5). .... 88

3.7 Longwave optical depth for wavelengths 4-100µm as a function of $q$, g/kg. Evaluated from SBDART data using Equation 3.13. ............... 90
3.8 Black-body spectrum for the peak SST value (300.15 K). Shaded area shows the wavelengths parameterised as in the window region.

3.9 Longwave optical depth as a function of $q$, g/kg, for wavelengths (left) 4-8 and 14-100$\mu$m and (right) 8-14$\mu$m calculated from SBDART data using Equation 3.13. Bottom panels zoom in on $q$ between 0 and 2 g/kg to show fit at lower humidity. Blue lines indicate the fit achieved by Equations 3.14 and 3.15.

3.10 Solid black line: Fraction of blackbody emission in the window region as a function of temperature, $T$. Dashed red line: Fit to this using Equation 3.16.

3.11 As Figure 3.9, but for the parameterisation in Equations 3.17 and 3.18.

3.12 (Left) longwave and (right) shortwave heating rates, K/day, for the reference profiles shown in Figure 3.1 based on the parameterisations presented in this chapter and used in the MITgcm (Equations 3.14, 3.15 and 3.4).

3.13 Shortwave heating rates, K/day, for the reference profiles shown in Figure 3.1 based on the parameterisation of Lacis and Hansen [1974].

3.14 Fractional difference from SBDART shortwave heating of (left) the new shortwave parameterisation and (right) that of Lacis and Hansen [1974].

3.15 Shortwave heating rates, K/day, for the alternative scheme parameterised using a polar humidity profile (Equation 3.5).

3.16 Longwave heating rates, K/day, for the alternative scheme allowing for varying blackbody emission by the window region (Equation 3.16).

3.17 Zonal mean zonal wind speed, m/s, (shading) and temperature, K, (contours) from the MITgcm when run with the radiation scheme presented here using the fixed longwave window fraction.

3.18 As Figure 3.17 but with the radiation scheme using the varying blackbody window fraction (Equation 3.16).

4.1 Zonal mean of total diabatic heating rates $Q$, K/day, for the experiments where humidity is varied. Contour interval is 0.5 K/day; grey line indicates zero contour.

4.2 Contours show the contribution to the heat budget of the sum of the eddy heat flux convergences $-\frac{\partial\theta'}{\partial \phi}$ and $-\frac{\partial\omega'}{\partial \phi}$. Contour interval is 0.5 K/day. Arrows indicate the eddy heat flux: $(\nu'\theta', \omega'\theta')$.

4.3 As Figure 4.1 but for radiative contributions to heating rates, K/day. Contour interval is 0.5 K/day.
4.4 As Figure 4.1 but for diffusive contributions to heating rates, K/day.  
Contour interval is 0.5 K/day. .................................................. 109

4.5 As Figure 4.1 but for convective contributions to heating rates, K/day.  
Contour interval is 0.5 K/day. .................................................. 111

4.6 As Figure 4.1 but for large-scale latent heat release contributions to heating rates, K/day. Contour interval is 0.5 K/day. .......................... 112

4.7 Zonal mean potential temperature, K, for the experiments where humidity is varied. ......................................................... 113

4.8 PDFs of the difference between the lapse rate and saturated adiabatic lapse rate, K/km, at 500 hPa for latitudes within 10° of the Equator. .. 114

4.9 PDFs of the difference between the lapse rate and saturated adiabatic lapse rate, K/km, in the storm track at 500 hPa. The storm track is approximated as a 30° latitude band centred over the latitude of peak \( \overline{v^2 T^0} \) (see Table 2.2) ......................................................... 116

4.10 Meridional temperature difference across the storm track vs bulk moist stability................................................................. 118

4.11 Zonal mean meridional mass transport in potential temperature coordinates, 10⁹kg/s. Red line shows the SST. .............................. 119

4.12 Shading shows the zonal mean zonal wind speed, m/s, contour interval 5ms⁻¹. Black contours show the zonal mean meridional mass transport, 10⁹kg/s. Note that contour interval for Hadley cell is 30(10⁹kg/s), while for Ferrel cell is 10(10⁹kg/s). ......................................................... 121

4.13 Left: Hadley cell heights predicted by Equation 1.34 compared with modelled height, with error bars indicating uncertainty on model values due to vertical resolution. Right: Hadley cell extents predicted by the theories in Section 1.3.2 compared with modelled extents. Black crosses indicate the latitude estimated from Equation 1.36. Green crosses are the latitude where the ratio \( (\overline{p_s} - \overline{p}_e)/(\overline{p_s} - \overline{p}_h) \) (see Equation 1.37) exceeds 1. The dotted line indicates a 1:1 match between model result and theory. Symbols identify individual experiments (see Table 2.1 or key on Figure 4.10 above). ......................................................... 122
4.14 Vertical integral of the terms in the angular momentum budget (Equation 1.40) for each run for levels higher than 660 hPa. Solid line represents $C_{\text{total}}$, dot-dashed line $C_{\text{zonal}}$, dashed line $C_{\text{eddy}}$, and blue line $f\bar{v}$. Thin lines indicate the latitude of the peaks of the upper level and surface zonal wind speeds. .................................................. 125

4.15 Eady growth rate, Equation 1.44, averaged from 900 to 260 hPa. .............. 127

4.16 Arrows show the Eliassen-Palm flux (Equations 1.24 and 1.25) for each experiment, scaled following Edmon et al. [1980]. See reference arrows in first panel for scale. Shading shows the E-P flux divergence. Boundary layer data is masked for each run for clarity. ................................. 128

4.17 Comparison of difference in specific humidity, g/kg, between (left) tropheat and 050 experiments and (right) 075 and 050 experiments. Contours show the specific humidity for the 050 experiment for reference. ... 132

4.18 Comparison of difference in potential temperature, K, between (left) tropheat and 050 experiments and (right) 075 and 050 experiments. Contours show the potential temperature for the 050 experiment for reference. 133

4.19 Zonal mean zonal wind speed, m/s, (shading) and zonal mean meridional mass transport, $10^9$kg/s, (contours) for the tropheat experiment. Note that contour interval for Hadley cell is $30(10^9$kg/s), while for Ferrel cell is $10(10^9$kg/s). ................................................................. 134

4.20 Comparison of difference in zonal mean zonal wind speed, m/s, between (left) tropheat and 050 experiments and (right) 075 and 050 experiments. Contours show the zonal wind speed for the 050 experiment for reference. 135

4.21 Comparison of difference in E-P flux (arrows) and E-P flux divergence (shading), between (left) tropheat and 050 experiments and (right) 075 and 050 experiments. See reference arrows in left hand panel for scale. ... 135

4.22 Wavenumber distribution as a function of latitude for each experiment, calculated as a 10-year average of Fourier transforms of $v'$ snapshots, and averaged over all pressure levels. The magenta line shows the Rossby deformation radius, while the white line shows Rhines scale. The white dashed line shows the latitude of the peak of the vertically averaged zonal wind speed, indicating the latitude where the jet is strongest. .......... 139

4.23 As Figure 4.22, but with the wavenumber normalised at all latitudes by $\cos(45)/\cos(\phi)$, indicating the equivalent physical scale at 45°. .......... 140
5.1 Instantaneous snapshots of the total precipitation rate, kgm$^{-2}$s$^{-1}$, equivalent to mm/s, for each experiment.

5.2 Shading shows snapshots of absolute vertical velocity, Pa/s, at 700 hPa. Blue indicates ascent, red descent. Contours show eddy geopotential height, m, at 700 hPa for comparison. Contour interval is 50m. Dashed lines indicated negative values, solid lines positive values.

5.3 Snapshots of eddy temperature, K, at 820 hPa.

5.4 Snapshots of $v'T'$, Kms$^{-1}$, at 820 hPa.

5.5 PDFs of (top) $\Phi'$ at 700 hPa, (middle) $T'$ at 820 hPa, and (bottom) $v'$ at 820 hPa for the dry (red) and wet (blue) simulations.

5.6 Composites of diabatic heating, $Q$, K/day (shading) and eddy temperature, K (contours). Composites are averages of variables centred on peaks of the product of relative vorticity and horizontal temperature gradient, $\zeta \nabla_H T$, for latitudes within 14° of the storm track latitude as defined in Table 2.2. Grey line marks the zero contour for the diabatic heating.

5.7 As Figure 5.6, but for the contributions of the different heating terms, K/day, to the total diabatic heating in the wet experiment.

5.8 As Figure 5.6, but for the vertical wind speed, Pa/s.

5.9 As Figure 5.6, but for the meridional wind speed, m/s.

5.10 Zonal and temporal mean eddy transport of (left) moist static energy and (right) sensible heat, averaged from surface to tropopause at the latitude of strongest $\overline{wT}$, as a function of reference $e_{0}$ fraction. The black line indicates the total, the red and blue show warm and cold sector contributions. Dashed lines in the right hand panel indicate estimated transport, see Section 5.2.3. Shaded areas indicate standard error, assuming data is uncorrelated after 10 days.

5.11 (left) As Figure 5.10 but for $\overline{T^2}$ (total not shown). (right) As before but for $\overline{v^2}$ and with red (blue) line indicating positive (negative) $v'$. Solid lines indicate pressure averages over the free troposphere, dashed lines averages over the boundary layer.

5.12 Eddy kinetic energy versus mean available potential energy for varying humidity experiments.

5.13 Ratio of EKE/MAPE as a function of $e_{0}$ fraction.
List of Tables

2.1 Summary of experiments performed with the MITgcm, including symbols
used for plotting data ............................................. 66
2.2 Latitude of peak $v'\theta'$ to be used to indicate storm-track position, and
tropopause and boundary layer height pressure levels at this latitude. .. 69
2.3 Summary and explanation of non-default values of input parameters used
in the SBDART runs. See Ricchiazzi [2002] for full list of parameters and
defaults. ............................................................ 76
1

Introduction

Water vapour is an important component of our climate system. Radiatively, it maintains the Earth at a temperature that can support life. Evaporation, transport, and precipitation of water vapour determine local climate, leading to warm, wet tropics, dry subtropics, and variable, stormy midlatitudes. When water vapour condenses, latent heat is released. This provides complex feedbacks onto the atmospheric circulation, the effects of which on static stability, the Hadley cell strength and width, and the development of eddies are still not well understood [Schneider et al., 2010]. Studies such as Frierson et al. [2006, 2007a] and O’Gorman and Schneider [2008a] have begun to use idealised aquaplanet model frameworks to explore the effects of latent heat feedbacks. Comparing the results of this kind of simple moist model with theories proposed after research with dry models, for example those using the physics of Held and Suarez [1994], allows new moist theories, more relevant to the real atmosphere, to
be investigated.

In this thesis, simulations run with an aquaplanet model, over a range of values of atmospheric moisture content, will be used to test theories for the controls on the circulation. The model used is the MITgcm, run with a simple physics package. In particular, the effects of latent heat release on the development of midlatitude eddies, and heat transport by these, will be examined. Previous studies on latent heat feedbacks have used very basic parameterisations of radiative transfer. Radiative processes redistribute the incoming solar energy, and the mean climate is sensitive to the radiative scheme used. A first step in the set-up of the aquaplanet simulations has therefore been to address this by creating a new parameterisation of the radiative effects of water vapour.

To give context to the features of the circulation to be investigated, this chapter will firstly give an overview of the climate observed on the kind of aquaplanet to be used in the experiments presented here, and of the equations governing the circulation. Following this, a summary of the essentials of radiative transfer will be given. Finally, the current understanding of the effects of latent heat on the atmospheric circulation will be reviewed.

1.1 Aquaplanets and the general circulation

With no topography or variations in surface type, aquaplanet experiments allow investigation of basic processes in geophysical fluid dynamics. The idealised set-up provides a simplified circulation, while still capturing many of the zonal mean features observed in the real world. Aquaplanets have been used for a range of purposes, for example investigations into the storm track using fixed sea surface temperatures (SSTs) [Brayshaw et al., 2008], or in an atmosphere-ocean coupled configuration for research into ocean circulation and the feedbacks of this onto the atmosphere [Marshall et al.,
2007, Enderton and Marshall, 2009, Ferreira et al., 2010].

1.1.1 Climate on an aquaplanet

‘The Aquaplanet Experiment’ (APE) [Neale and Hoskins, 2001] was designed for intercomparison of the physical parameterisations of climate models. A range of models with different physics schemes were run with fixed SSTs. The results provide a useful climatology of the atmospheric circulation on an aquaplanet. As an example of the kind of climate observed in this idealised set-up, Figure 1.1 shows the zonal mean zonal wind speed, temperature, meridional wind speed, pressure velocity, specific humidity, and relative humidity from the multi-model mean of the APE Atlas [Williamson and Coauthors, 2012], which presents the results of the experiment.

The zonal mean climatology shown in Figure 1.1 is not dissimilar to that of the Earth. A strong zonal jet stream is found at upper levels. Air rises over the equator and diverges towards the poles, before descending in the subtropics. These circulations are driven by, and in balance with, an equator to pole temperature gradient. In the midlatitudes, the meridional temperature gradients and vertical wind shear result in baroclinic instability, and in this region waves are generated which continue the poleward transport of momentum, heat, and humidity to higher latitudes.

The meridional circulation also determines the distribution of water vapour in the atmosphere. The lower central and right hand panels of Figure 1.1 show the modelled specific and relative humidities. The maximum evaporation occurs in the subtropics, where the air is relatively dry. Moisture is advected equatorward by the lower branch of the Hadley Cell towards the Intertropical Convergence Zone. Here near-saturated air is transported upwards by the circulation, resulting in cooling so that water precipitates out. The air then descends and warms in the subtropics, so that the saturation vapour pressure of an air parcel increases and relative humidity decreases. In the
Figure 1.1: APE multi-model means. Top row: zonal average of zonal wind speed, m/s, temperature, K, and meridional wind speed, m/s. Bottom row: zonal average of vertical wind speed, mb/day, specific humidity, g/kg, and relative humidity. Adapted from Williamson and Coauthors [2012] (Figure 4.15).
near-saturated tropics, latent heat release will affect the vertical temperature gradient, and feed back onto the height and strength of the Hadley Cell. In the midlatitudes, moisture carried by weather systems precipitates along fronts. This source of diabatic heating results in complex interactions between eddies and moisture. To understand how the different features of the circulation shown in Figure 1.1 interrelate it is useful to look at the governing equations.

1.1.2 Equations determining the circulation

Assuming hydrostatic equilibrium, the equations governing the dynamics of the atmosphere can be expressed in Cartesian and pressure coordinates as:

\[
\begin{align*}
\frac{D u}{D t} - f v + \frac{\partial \Phi}{\partial x} &= \mathcal{F}^{(x)} \quad (1.1) \\
\frac{D v}{D t} + f u + \frac{\partial \Phi}{\partial y} &= \mathcal{F}^{(y)} \quad (1.2) \\
\frac{\partial \Phi}{\partial p} &= -\frac{1}{\rho} \quad (1.3) \\
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} &= 0 \quad (1.4) \\
p &= \rho R_a T \quad (1.5) \\
\frac{D \theta}{D t} &= Q \quad (1.6) \\
\frac{D q}{D t} &= E - P \quad (1.7)
\end{align*}
\]

where the symbols used are described in the list of commonly used symbols on Page 11, and \( \frac{D}{D t} \) is the Lagrangian derivative, the change in a given quantity following a fluid parcel:

\[
\frac{D}{D t} = \frac{\partial}{\partial t} + u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + \omega \frac{\partial}{\partial p} \quad (1.8)
\]

Equations 1.1 and 1.2 are the horizontal components of the Navier Stokes equation,
describing horizontal momentum conservation. Equations 1.3 to 1.7 describe hydrostatic equilibrium, conservation of mass, the ideal gas law, conservation of heat, and conservation of water vapour respectively.

Useful diagnostic equations may be derived from Equations 1.1-1.7. Taking the dominant terms of Equations 1.1 and 1.2 gives geostrophic balance:

\[ f v \approx \frac{\partial \Phi}{\partial x} \]  
\[ f u \approx -\frac{\partial \Phi}{\partial y} \]  

Combining this with the ideal gas law gives the thermal windshear equations:

\[ f \frac{\partial v}{\partial \rho} \approx -\frac{1}{T \rho} \frac{\partial T}{\partial x} \]  
\[ f \frac{\partial u}{\partial \rho} \approx \frac{1}{T \rho} \frac{\partial T}{\partial y} \]

These allow some estimation of how changes in atmospheric heating will be reflected by the circulation. For example, increasing the meridional temperature gradient can be expected to result in shifts and changes to the strength of the zonal jet. This balance, indicated by Equation 1.12, is illustrated by comparing the zonal wind and temperatures in Figure 1.1, where it can be seen that stronger meridional temperature gradients are associated with stronger vertical gradients of zonal wind speed.

The zonal mean climate shown in Figure 1.1 does not give the full picture of the circulation. Particularly in the midlatitudes, baroclinic instability due to the meridional temperature gradient results in large deviations from the mean state. A common approach to understanding the general circulation is therefore to divide quantities into some mean state, either zonal or temporal, and then to look at deviations, or eddies,
around this state. Defining eddies as deviations from the zonal mean, a given field could be expressed as:

\[ X = \overline{X} + X' \]  

(1.13)

where the over-line indicates the zonal mean, and the dash a deviation from this. Alternatively, the eddies may be viewed as deviations from a temporal mean:

\[ X = [X] + X^* \]  

(1.14)

where square brackets denote the temporal mean, and stars refer to deviations from this mean. A given field could then be expressed [Peixoto and Oort, 1992]:

\[ X = [X] + [X]' + \overline{X}' + X'^* \]  

(1.15)

In Chapter 4, the zonal and temporal mean of eddy fluxes, which involve products of two fields, will be considered. Expanding the zonal and temporal mean of a given field product, for example \( uv \), into the above mean state and eddy contributions gives:

\[ \overline{uv} = [\overline{u}][\overline{v}] + [u'][v] + [\overline{u}'v] + [u'^*v'^*] \]  

(1.16)

where the terms on the right hand side are the momentum fluxes associated with the zonal and temporal mean state, stationary eddies, transient zonal mean circulations, and transient asymmetric eddies respectively. On an aquaplanet, which is zonally symmetric, the time mean at a given location is similar to the zonal mean, and the final term is expected to dominate in the mid-latitudes. In Chapter 4, plots of eddy quantities include all eddy terms, and are evaluated as \( \overline{uv} - [\overline{u}][\overline{v}] \). For brevity, these are denoted as for example, \( \overline{u'v'} \). In Chapter 5, to give an indication of the
The eddy-mean flow perspective discussed above can be applied to the governing equations to give useful insight into the mechanisms of momentum and heat transport throughout the atmosphere. Multiplying Equation 1.4 by $u$, adding this to Equation 1.1, separating variables into zonal mean and eddy quantities, and approximating vertical momentum advection as negligible gives:

$$\frac{\partial \pi}{\partial t} = \mathcal{F} \left( f - \frac{\partial \pi}{\partial y} \right) - \frac{\partial \bar{w} \bar{v}'}{\partial y} + \mathcal{F}^{(x)}$$  \hspace{1cm} (1.17)

The zonal mean zonal wind speed can thus be interpreted as controlled by a balance of advection of zonal mean planetary vorticity ($f \mathcal{F}$) and relative vorticity ($\pi (\pi')_{\pi'}$), convergence of eddy angular momentum ($\frac{\partial \bar{w} \bar{v}'}{\partial y}$), and friction ($\mathcal{F}^{(x)}$). In the midlatitudes, $\pi (\pi')_{\pi'}$ is small compared to the other terms in the budget, and is often approximated to be negligible. The feedbacks from the eddies onto the mean state help to drive the zonal jet. Where wave activity is stronger at higher latitudes, these may result in a second, entirely eddy driven, jet, separate from the subtropical Coriolis driven component of the jet.

A similar approach can also be used to derive an equivalent equation for the zonal mean heat budget in terms of eddies:

$$\frac{\partial \theta}{\partial t} = -\mathcal{F} \frac{\partial \theta}{\partial y} - \frac{\partial \bar{w} \bar{v}'}{\partial p} - \frac{\partial \bar{v} \bar{v}'}{\partial y} - \frac{\partial \bar{w} \bar{v}'}{\partial p} + \overline{Q}$$ \hspace{1cm} (1.18)

In quasi-geostrophic scaling, the terms describing advection by the mean meridional circulation and vertical eddy flux divergence are comparatively small, and are often neglected [Holton, 2004].* There is then a strong cancellation between the second

*NB. Figure 4.2, Chapter 4, demonstrates that $-\mathcal{F} \frac{\partial \bar{w}}{\partial p}$ does in fact contribute to the heat budget,
and third terms on the right hand side of Equation 1.18, corresponding to adiabatic cooling by ascending motions and meridional eddy heat flux divergence. Making this quasi-geostrophic approximation, in order to separate the component of the meridional circulation that balances diabatic heating, Andrews and McIntyre [1976] introduced the transformed Eulerian mean (TEM) formulation of the primitive equations. They defined a residual meridional circulation as:

\[
\bar{v}^* = \bar{v} - \frac{\partial \bar{v}' \bar{\theta}' / \partial p}{\partial p} \\
\bar{w}^* = \bar{w} + \frac{\partial \bar{v}' \bar{\theta}' / \partial y}{\partial y} \\
\frac{\partial \bar{v}^*}{\partial y} + \frac{\partial \bar{w}^*}{\partial p} = 0
\]

(1.19)  (1.20)  (1.21)

where \( \bar{\theta}_p \) denotes the derivative of \( \theta \) with respect to \( p \). Note that asterisks here denote the TEM circulation, rather than temporal eddies as above. This allows Equations 1.17 and 1.18 to be re-expressed as:

\[
\frac{\partial \bar{v}}{\partial t} = f \bar{v}^* + \mathcal{F}(x) + \nabla . F \\
\frac{\partial \bar{\theta}}{\partial t} = \bar{Q} - \bar{v}^* \frac{\partial \bar{\theta}}{\partial p}
\]

(1.22)  (1.23)

This formulation describes a residual meridional circulation driven by diabatic heating and by eddy flux convergence. In Equation 1.22, \( \nabla . F \) is the divergence of the Eliassen-Palm (E-P) flux, which provides a way to visualise the propagation of waves, and is defined in the quasi-geostrophic approximation and Cartesian coordinates as: with a net flux of heat from lower to upper levels in the midlatitudes.
\begin{align*}
F_{(y)} & = -\overline{u'u'}^l \\
F_{(p)} & = \int \overline{v'\theta'/\theta_p} \end{align*} \quad (1.24)

Edmon et al. [1980] give an overview of the uses of the E-P flux. The flux itself can be taken as an indication of the direction and strength of Rossby wave propagation in the \((y, p)\) plane, provided the eddy activity can be considered wavelike. Its divergence indicates the strength of the forcing of the zonal mean state by eddies (see Equation 1.22), regardless of whether they can be considered as propagating or turbulent.

1.2 Radiative Processes

Equations 1.18 and 1.23 show the importance of diabatic heating, \(Q\), in shaping the zonal mean circulation. While the key aim of this thesis is an investigation of latent heat feedbacks, radiative heating and cooling are important in determining atmospheric structure. A first strand of research involved the development of a radiation parameterisation for the model (Chapter 3) to improve the mean climate against which experiments could be carried out. This section provides background on the relevance of radiative heating and the processes involved.

1.2.1 Radiation and Climate

Absorption and emission of radiation have a strong influence on the structure of our climate. The distributions of incoming solar radiation and outgoing longwave radiation are such that net energy is gained at low latitudes, and lost at high latitudes, as shown in Figure 1.2. The equator-to-pole contrasts in these distributions require heat to be transported poleward by atmospheric circulations.
**Figure 1.2:** Zonal average of absorbed shortwave radiation (ASR) and outgoing longwave radiation (OLR) as a function of latitude, with the shading indicating the surplus in the tropics and deficit at high latitudes. Data is from the CERES experiment [Wielicki et al., 1996]. Adapted from Donohoe and Battisti [2012] (Figure 1). ©American Meteorological Society. Used with permission.

In addition to driving the large-scale circulation, radiative processes are also important in determining the vertical temperature structure. In the troposphere, radiative processes tend to produce convective instability so that heat is redistributed by vertical motions up to a height where radiative heating by ozone produces the very stable stratosphere. This provides a constraint on the height to which the Hadley cell extends, and will be discussed in further detail in Section 1.3.2.

Radiatively, water vapour is an important gas in the atmosphere, being the largest contributor to the natural greenhouse effect. As the atmosphere warms under climate change, the amount of water vapour it holds increases at around 7% per degree, providing a powerful radiative feedback [Myhre et al., 2013]. Water vapour is radiatively active at both short and long wavelengths. Figure 1.3 shows the absorptance of six gases, including water vapour, as a function of wavelength. It can be seen that water vapour absorption has a complex wavelength dependence, varying from very strongly to very weakly absorbing. One particularly notable feature of Figure 1.3, when all gases
Figure 1.3: Clear sky absorption spectra for common, strongly absorbing gases in the atmosphere. Taken from Figure 3.14, Chapter 3 of Andrews [2010]. ©Cambridge University Press 2000.
are considered (lowest panel), is the region of weaker absorption by the atmosphere between approximately 8 and 14 μm. This is known as the ‘infrared window’ region. It is important that the low absorption in this region receives appropriate treatment alongside the more strongly absorbing regions in the parameterisation of longwave heating rates in the atmosphere. As shown in Figure 1.1 (lower middle panel) unlike carbon dioxide, water vapour is not well mixed spatially, with a distribution dependent on air temperature and circulation patterns. Inclusion of its effects in a model therefore requires parameterisations that can describe the variation of absorption with wavelength and that can respond to local humidity values.

1.2.2 Radiative transfer

As a beam of radiation passes through the atmosphere, each wavelength can interact with the gases and aerosols present via absorption, emission, and scattering. The relative importance of these processes is both wavelength and temperature dependent. At the temperatures observed in the atmosphere, emission is negligible at shorter wavelengths but a key process in the longwave region. Scattering however is most significant in the shortwave. Absorption occurs throughout the spectrum, but its magnitude is again wavelength dependent, as shown in Figure 1.3.

The equation of radiative transfer may be expressed [Liou, 2002]:

\[ dI_\lambda = -k_\lambda \rho (I_\lambda - J_\lambda) ds \]  \hspace{1cm} (1.26)

In the above, \( I_\lambda \) and \( J_\lambda \) are the intensity of incident radiation and the radiative source function at a given wavelength, \( \lambda \). The extinction coefficient \( k_\lambda \) describes the attenuation of the beam by the combination of absorption and scattering of radiation. This equation therefore states that over a path of length \( ds \), the change in intensity
of an incident beam of radiation at a given wavelength will be equal to the amount
gained through emission or scattering from elsewhere into the beam \( k_{\lambda} \rho J_{\lambda} \) less the
amount lost over the path \( k_{\lambda} \rho I_{\lambda} \).

The extinction coefficient for a given absorber is dependent on its molecular struc-
ture and shape. The molecular structure determines the rotational, vibrational, and
electronic energy levels of the material. The molecule may absorb or emit an incident
photon of an appropriate wavelength to alter its energy state. These lines may addi-
tionally be broadened by collision between molecules, or by the Doppler effect, as
molecules may have different velocities. The size and shape of the absorber determine
its scattering properties. If a particle is much smaller than the wavelength of the ra-
diation, it may be polarised by the electric field of the light, becoming an oscillating
dipole. The radiation produced is known as Rayleigh scattering. For a particle of com-
parable size to, or larger than, the wavelength of the radiation, the Rayleigh scattering
model breaks down. In this case Mie theory is used, which approaches the problem by
solving the electromagnetic wave equation for interaction with a homogenous sphere.

It can be seen that interaction of radiation with the constituents of the atmosphere is
not a simple problem. The most accurate calculation of radiative transfer is therefore
to evaluate the transmission over each absorption line over all angles of incidence.
However, this makes calculating transmission over the entire spectral range, at each
level of the atmosphere, computationally intensive. Alternatively, the spectrum can
be divided into regions containing large numbers of lines, and approximations made to
compensate for not making the full calculation.

Further useful simplifications to radiative transfer are the two stream, grey at-
mosphere, and plane parallel approximations. The first of these approximates light
as travelling in only two discrete directions, averaging scattered radiation over angle.
The result is two coupled differential equations for the upward and downward beams,
which may be solved for given absorption and scattering coefficients and source functions [Meador and Weaver, 1980]. To remove the complication of considering spectral lines or bands in detail, the grey atmosphere approximation assumes that the absorption coefficient is constant over wavelength. Finally, if the thickness of the atmosphere is much less than the radius of the planet, it can be approximated as plane parallel, so that the effects of curvature are ignored.

In an intermediate complexity general circulation model (GCM) such as that used here, all physics processes are parameterised using simple schemes, allowing a range of idealised experiments to be performed by varying constants. A simpler radiation scheme is therefore better suited for this model. However, it must still provide a sensible description of the processes involved and produce a useful climatology to base experiments around, comparable with other studies [e.g. Williamson and Coauthors, 2012] and the real atmosphere. Current parameterisations used in the model will be described in Section 2.4.2. In Chapter 3, the development of an improved scheme will be presented.

1.3 Latent Heat Release and Zonal Mean Climate

Alongside radiative processes, the other major source of diabatic heating is the release of latent heat. The current understanding of the role of latent heat release in climate will now be reviewed.

1.3.1 Static Stability

Latent heat release has long been known to affect the static stability of the atmosphere [Stone and Carlson, 1979, Emanuel, 1988]. As can be seen from Equation 1.12, changes to the thermal structure of the atmosphere are balanced by changes to the circulation, with effects on both the mean state and on eddy development. A theory for how the
static stability is determined is therefore important in understanding the effects of latent heating.

The Clausius-Clapeyron equation describes the transition between two phases of matter. For the transition between water vapour and liquid water in the atmosphere, this can be written:

\[
\frac{de_s}{dT} = \frac{L e_s}{R_v T^2} 
\]  

(1.27)

showing that the saturation vapour pressure, \(e_s\), of a parcel of air decreases as temperature decreases. An ascending air parcel cools adiabatically, so that at some point (the lifting condensation level) this parcel will become saturated, and latent heat will be released. The unsaturated parcel, with no latent heat release, behaves similarly to a dry parcel, following the dry adiabatic lapse rate, \(\Gamma_d = g/c_p\). However, for the saturated parcel, latent heat release will alter the stability. By consideration of the heat budget of the ascending saturated parcel, the saturated adiabatic lapse rate, \(\Gamma_s\), may be derived:

\[
\Gamma_s = -\frac{dT}{dz} = \frac{g}{c_p} \left(1 + \frac{L \mu_s}{R_v T}\right) \left(1 + \frac{L^2 \mu_s}{c_p R_v T^2}\right) 
\]

(1.28)

The ascending parcel releases heat to the atmosphere, which warms so that the saturated adiabatic lapse rate is more stable than dry neutrality. The static stability of an air parcel is therefore dependent on its moisture content. If the atmospheric lapse rate is below moist neutral, the parcel will be absolutely stable. If it is above dry neutral, air parcels will tend to quickly rearrange until a stable profile is reached. Between these two limits is conditional instability, where the stability of a parcel depends on whether or not it is saturated.
Figure 1.4: Meridional temperature difference across the storm track (defined to be between 20° and 70°N at 500mb) vs bulk moist stability at 50°N for NCEP reanalysis data. Adapted from Juckes [2000] (Figure 6). ©American Meteorological Society. Used with permission.

While the static stability of a given air parcel moving in a given background atmospheric profile may be understood in this way, how this then relates to the mean stability of the atmosphere as a whole is more complex. In the tropics, where the air is near saturation, and eddy activity is limited, the troposphere is generally observed to be close to moist neutrality, and static stability is controlled by convective processes [Stone and Carlson, 1979, Emanuel et al., 1994].

In the midlatitudes however, despite the reduced moisture compared with the tropics, the atmosphere tends to be even more stable than moist neutrality [Stone and Carlson, 1979]. This is generally understood to be related to the presence of baroclinic eddies, which transport heat vertically, with a range of theories attempting to explain their role in determining the stratification, and how moisture can be expected to affect this.

One school of thought that has developed is based on a theory proposed by Juckes [2000]. This attempts to relate differences between equivalent potential temperature,
\( \theta_{eq} \), which he defines as:

\[
\theta_{eq} = \left( T + \frac{Lq}{c_p} \right) \left( \frac{p_0}{p} \right) \frac{\tilde{\eta}}{\tilde{\eta}}
\]

at the surface and tropopause to the midlatitude temperature gradient using a series of simple assumptions. Firstly, it is assumed that the equivalent potential temperature at the surface is roughly equal to the potential temperature at the tropopause, so that the minimum difference between the two is zero, i.e., moist neutrality. This is consistent with the observation of near-neutral moist stability in cyclones [Emanuel, 1988, Korty and Schneider, 2007, Pauluis et al., 2008], and relies on the idea that this lower bound on static stability is determined convectively. The mean difference is then assumed to be equal to the minimum (zero) plus half the variance. Finally, the variance in the storm track is assumed to arise from adiabatic meridional advection by transient eddies, and thus be proportional to the temperature gradient across the track. The result is a proportionality between surface-tropopause equivalent potential temperature difference and midlatitude temperature gradient:

\[
\Delta_z \bar{\theta}_{eq} \propto \Delta_y \bar{T}
\]

Figure 1.4, taken from Juckes [2000], shows an evaluation of this theory for NCEP reanalysis data, using the bulk stability at 50°N, and the temperature difference between 20° and 70°N at 500mb. Qualitatively the results appear consistent with the suggested linear relationship with zero intercept. This approach is appealing in its simple theoretical basis, compared with attempts to adapt dry theories [e.g., Held, 1982, Schneider, 2004]. Later studies found that this theory could account for the lapse rates observed in a simplified climate model under a range of moisture contents [Frierson et al., 2006, Frierson, 2008], lending support to the hypothesis.
Figure 1.5: Vertical cross sections of (a) the zonal wind speed and (b) the mean meridional streamfunction, taken from annual-mean NCEP-NCAR reanalysis data between 1979 and 2001. Dashed contours indicate negative values, heavy contour shows the zero line for the zonal wind. (a) Contour interval is 5 m/s for positive and 2 m/s for negative values. (b) Contour interval is $2 \times 10^{10}$ kgs$^{-1}$, with only odd contours plotted. Taken from Dima et al. [2005] (Figure 1). ©American Meteorological Society. Used with permission.

In contradiction to this result, Schneider and O’Gorman [2008] found that altering the lapse rate used in the convection scheme in simple model experiments did not appear to strongly control the extra-tropical stability. Varying the optical depths in their model, they found that the theory of Juckes [2000] failed to account for colder, drier climates, which were better explained with dry theories [Schneider, 2004]. Their results suggest that bulk moist stability may not always clearly relate to the gradient of equivalent potential temperature. Equation 1.30, and the validity of the underlying assumptions involved, will be examined in Chapter 4.

1.3.2 THE HADLEY CELL

A dominant feature of the atmospheric circulation is the Hadley cell. This is strongly linked to the water cycle in the tropics and subtropics, so that changes to its width will
significantly alter the climate at a given latitude, making the difference between moist tropical rainforest or dry subtropical desert. The cell extent is also closely related to the positions of the zonal jets and storm tracks, as the meridional overturning is a key element in the momentum budget (e.g. Equation 1.17). Current theories suggest that the height and width of the cell are related to static stability and meridional temperature gradient, so that changes to these will lead to changes to the cell, but the expected effects of increased moisture are unclear [Schneider et al., 2010]. The zonal mean meridional overturning can be visualised by defining a mass streamfunction, \( \Psi \):

\[
\bar{v} = \frac{g}{\partial p} \Psi
\]

(1.31)

\[
\bar{w} = -\frac{g}{\partial y} \Psi
\]

(1.32)

This streamfunction can be seen to satisfy mass conservation, Equation 1.4, in the zonal mean. In isobaric and spherical coordinates, the streamfunction may be calculated from \( v \):

\[
\bar{\Psi} = \frac{2\pi acos\phi}{g} \int_0^\theta \bar{v} dp'
\]

(1.33)

An example of the streamfunction and zonal jet structures observed in reanalysis data is shown in Figure 1.5, taken from Dima et al. [2005]. Air ascends to approximately 100 hPa in the tropics, travels polewards aloft, descends, and returns equatorward, with the Hadley cells extending to roughly 30\(^\circ\) in the Northern and Southern hemispheres. The upper branch of the cells are in balance with zonal jets, as indicated by Equation 1.17. In the midlatitudes, weaker, eddy-driven Ferrel cells are found, overturning in
the opposite sense. Comparing with Figure 1.1 demonstrates the ability of aquaplanet simulations to capture the key features of the circulation.

The Hadley cell height is relatively well predicted through constraints on the tropical lapse rate (moist stability, see above) and radiative constraints. Based on the definitions of troposphere and stratosphere in terms of heat transport, the tropopause can be evaluated as the height above which radiative equilibrium is achieved, and below which entropy must be redistributed convectively [Held, 1982, Thuburn and Craig, 1997]. For a given lapse rate, the tropopause height can be then taken to depend on the surface temperature, and on a radiative constraint, so that the height is the minimum at which the temperature profile given by this lapse rate matches a radiative equilibrium temperature profile above this height. The tropopause height, $H_t$, will therefore vary with lapse rate, surface temperature, and atmospheric composition (as this will determine radiative emission at each level), and can be estimated as [Schneider, 2007]:

$$H_t \approx (1 - c) \frac{T_s}{\Gamma} + cH_e$$

(1.34)

where $c = 2^{-1/4}$, $T_s$ is the surface temperature, $\Gamma$ is the lapse rate, and $H_e$ is the emission height (where the atmospheric temperature is equal to the emission temperature).

For a given cell height, the factors controlling cell strength and width are less clear. From Equation 1.17, it can be seen that at upper levels, where the effects of friction may be neglected, the steady state zonal momentum balance is determined by advection of planetary vorticity, zonal mean relative vorticity, and eddy angular momentum flux divergence. Assuming the latter of these to be negligible, that the surface zonal wind is zero, and that the cell is energetically closed and confined to small angles, an analytical
expression for the width of the cell may be derived [Held and Hou, 1980]:

\[
\phi_{HH} \approx \left( \frac{5 g H_1 \Delta_H}{3 \Omega^2 a^2} \right)^{\frac{1}{2}}
\]

(1.35)

In the above, \( \Delta_H \) is the vertically averaged fractional change in potential temperature from equator to pole in radiative equilibrium. The cell strength (i.e. \( \tau \) in Equation 1.17) is then determined by thermal, rather than momentum, balance.

However, typically in both GCMs and data these approximations do not apply, with eddy angular momentum fluxes not negligible [Dima et al., 2005, Walker and Schneider, 2006]. Figure 1.6, adapted from Dima et al. [2005], shows evaluation of the terms in the angular momentum budget (Equation 1.17) for NCEP-NCAR reanalysis.
data. Comparing the upper and lower panels, showing the contributions of the mean meridional circulation (panel a) and eddy momentum flux convergence (panel b), it can be seen that these are of similar magnitude. The effects of these eddies on the Hadley cell is difficult to predict. One theory is that the cell could be approximated to terminate at the latitude where the wind shear of the jet predicted by the Held and Hou [1980] model becomes baroclinically unstable [Held and Fellows, 2000, Lu et al., 2007]. With this modification, the extent can be estimated as [Frierson et al., 2007b]:

\[ \phi_H \approx (H_t \Delta_v)^{1/4} \]  

(1.36)

where \( \Delta_v \) is the dry bulk stability (difference in potential temperature between surface and tropopause). Testing of this scaling has shown that it appears successful over a range of SST profiles [Frierson et al., 2007b].

Considering the limit where the cell extent is instead driven predominantly by eddy angular momentum flux divergence, Korty and Schneider [2008] suggest that the Hadley cell can be taken to terminate at the latitude where vertical wave fluxes become deep enough to reach the upper troposphere. This theory is justified by the idea that, at this latitude, the eddy momentum flux divergence term in the momentum budget is expected to change sign. Beginning from zonal momentum balance, Schneider and Walker [2006] estimated the pressure to which eddies redistribute entropy, \( p_e \), relative to the surface pressure as:

\[ p_0 - p_e = \frac{f}{\beta} \frac{\partial \bar{\theta}_s}{\partial \bar{\theta}} \]  

(1.37)

where \( \bar{\theta}_s \) is the mean near-surface potential temperature and \( \partial \bar{\theta}/\partial \bar{\theta} \) is the near surface static stability. Korty and Schneider [2008] therefore estimate the cell width as the
latitude at which the ratio $(\bar{p}_0 - \bar{p}_e)/(\bar{p}_0 - \bar{p}_t)$, where $p_t$ is the tropopause pressure, first exceeds a critical O(1) value. However, while this method gives a good measure of the pressure range of vertical wave activity in dry models [Schneider and Walker, 2006], this does not appear to extend to wet models [Schneider and O’Gorman, 2008].

From the theories summarised above, some predictions of the effects of latent heat release on the cell in the MITgcm can be made. As humidity increases, the lapse rate is expected to decrease, corresponding to an increase in dry bulk stability. Equation 1.34 suggests that the cell will deepen in response to this. With the SST fixed and with more latent heat release occurring in the tropics, the midlatitude temperature gradient is likely to increase. Equations 1.35 and 1.36 would therefore suggest a widening of the Hadley cell, while the prediction of Equation 1.37 depends on how both the stability and temperature gradient covary. It is more difficult to predict how the strength of the overturning will change. Both thermal contrasts and changes to baroclinicity may affect the momentum balance. With increasing meridional temperature gradient, a strengthening of the cell seems plausible.

1.3.3 The Jets and the Zonal Momentum Budget

As discussed in the previous section, changes to the distribution of heating in the atmosphere alter the angular momentum budget. As a result, the zonal jet may shift in latitude, intensify or weaken, and potentially divide into multiple jets [Lee and Kim, 2003, Son and Lee, 2005, Lorenz and DeWeaver, 2007]. The shear of the jet is strongly associated with the development of atmospheric waves (discussed below), which themselves act to maintain the jet (e.g. Equation 1.17). Shifts in the jet and storm tracks are therefore intimately linked. The jet position is sensitive to both tropopause height and meridional temperature gradient [Haigh et al., 2005, Son and Lee, 2005, Lorenz and DeWeaver, 2007]. Applying heating perturbations at different
pressures and latitudes in a simple GCM, Lorenz and DeWeaver [2007] found the jet to shift poleward (equatorward) as the tropopause height increased (decreased), or, for a heating perturbation applied in the troposphere, as the meridional temperature gradient increased (decreased). These shifts in jet latitude are consistent with the theories of the Hadley cell discussed above, with increases in meridional temperature gradient and cell height expected to result in increases in the width of the cell.

Zonal jets are generally divided into two types, depending on the dominant term in the zonal momentum budget, Equation 1.17. The subtropical jet is driven by the zonal mean angular momentum transfer in the Hadley cell [e.g. Held and Hou, 1980], while the eddy-driven jet is driven by convergence of eddy momentum in the midlatitudes [e.g. Williams, 1979, Panetta and Held, 1988]. Where the maximum baroclinicity is in the region of the subtropical jet, a single jet is observed. If the baroclinic zone is broader, an additional, separate eddy-driven component may be found [Lee and Kim, 2003, Son and Lee, 2005]. Changes to both the zonal mean and eddy components of the angular momentum budget have been shown to be needed in order to account for the shift in the jets when heating perturbations are introduced [Kushner and Polvani, 2004, Haigh et al., 2005, Simpson et al., 2009].

In the MITgcm simulations presented in this thesis, to allow the primitive equations to be integrated on the cube-sphere grid, vector-invariant forms of the momentum equations are used. Adding and subtracting (so that the sum is zero) \( v \frac{\partial}{\partial x} \) and \( \omega \frac{\partial}{\partial x} \omega \) to Equation 1.1, and rearranging, the zonal momentum equation can be expressed:

\[
\frac{\partial u}{\partial t} = f v + \zeta_3 v - \zeta_2 \omega - \frac{\partial B}{\partial x} + F^{(u)}
\]

(1.38)

In the above, \( \zeta_{2,3} \) are components of the vorticity vector: \( \zeta = \nabla \times v \), and B is the Bernoulli function: \( B = \Phi + KE \). In the zonal mean the zonal gradients are zero, and
this reduces to:

\[
\frac{\partial \bar{u}}{\partial t} = (f - \frac{\partial u}{\partial y})v - \frac{\partial u}{\partial p} \omega + \bar{F}(x)
\]  

(1.39)

Linearising about the zonal mean and integrating over pressure gives [cf. Haigh et al., 2005, Simpson et al., 2009]:

\[
\frac{1}{g} \frac{\partial}{\partial t} \int \bar{u} dp = C_{zonal} + C_{eddy} + \frac{1}{g} \int f \bar{v} dp + \frac{1}{g} \int \bar{F}(x) dp
\]

(1.40)

\[
C_{zonal} = -\frac{1}{g} \int \frac{\partial u}{\partial y} v + \frac{\partial u}{\partial p} \omega dp
\]

(1.41)

\[
C_{eddy} = -\frac{1}{g} \int \frac{\partial u'}{\partial y} v' + \frac{\partial u'}{\partial p} w' dp
\]

(1.42)

\[
C_{total} = C_{zonal} + C_{eddy}
\]

(1.43)

If the integral is performed over the whole atmosphere, mass conservation means that the term in \( f \bar{v} \) is negligible. The integral of the frictional term is then equal to the surface stress, \( \bar{\tau}_s \). If instead the integral is performed over only upper levels of the atmosphere, friction does not contribute, and the balance is instead between the advective terms. In steady state, the temporal mean of the left hand side of Equation 1.40 will be zero, so that the terms on the right hand side must be in balance.

1.4 Latent heat release and baroclinic eddies

In addition to the static instability that occurs when a column of fluid is heated from below or cooled from above, in a rotating fluid more complex types of instability may occur. From the above discussions of the controls of stability, the Hadley cell, and the momentum budget, it has been seen that the resulting eddies play an important role in the atmospheric circulation. Of particular relevance is baroclinic instability, which is
related to the vertical gradient of the zonal wind speed and the horizontal temperature gradients that balance this (Equation 1.12), and barotropic instability, which is related to strong horizontal gradients in the zonal wind speed [Andrews, 2010].

1.4.1 GROWTH AND DECAY OF STORMS

In the midlatitudes, where there are strong horizontal temperature gradients, baroclinic instability can lead to the generation of eddies, which transport heat and momentum, feeding back onto the zonal mean flow. Eddy development can be described theoretically by Eady’s model of baroclinic instability [Eady, 1949]. Taking a background flow on an f-plane that is a linear function of height between rigid boundaries, unstable modes can be shown to exist (see Andrews [2010] pp. 143-146). Based on this model, a useful diagnostic to give a measure of the baroclinicity of a climate system can be derived: the maximum Eady growth rate, $\sigma_E$. This can be expressed as [Brayshaw et al., 2008]:

$$\sigma_E = \frac{0.31 f \partial \pi}{N} = - \frac{0.31 g \partial \theta}{N \partial y}$$

(1.44)

$$N = \sqrt{\frac{g}{\partial \theta \partial z}}$$

(1.45)

where $N$ is the Brunt-Väisälä frequency, the oscillation frequency of an air parcel in a statically stable environment. The rate of development of baroclinic eddies is therefore related to the vertical shear of the zonal jet and to the static stability. The Eady growth rate can be used to estimate the latitude range over which eddies are likely to be generated.

A range of simple conceptual models of the development of storms exist [Bjerknes
Figure 1.7: Horizontal composites of the development of a midlatitude storm at (top to bottom) 300, 700, and 925 hPa. Data is from ERA-Interim reanalysis for DJF of 1989-2009. Left to right, columns show 48h and 24h before maximum intensity, and maximum intensity. Solid contours indicate geopotential height, m, dashed lines equivalent potential temperature, K, and arrows wind vectors relative to the system. Shading in the middle row indicates vertical velocity, Pa/s, and in the top row divergence. Taken from Dacre et al. [2012] (Figure 4). ©American Meteorological Society. Used with permission.
and Solberg, 1922, Shapiro and Keyser, 1990, Browning and Roberts, 1994], which discuss storm structure in terms of warm and cold fronts, upper and lower level waves, and motion of air masses relative to the system. By tracking and compositing storms in ERA-Interim data using methods developed by Catto et al. [2010], Dacre et al. [2012] generated an atlas of extratropical cyclones. These observations are consistent with conceptual models, and provide useful insight into cyclone development. Figure 1.7 shows the atlas results for the evolution of the vertical cyclone structure.

An upper level geopotential trough, associated with positive potential vorticity, moving over a region of baroclinicity may induce a circulation at lower levels [Hoskins et al., 1985]\(^1\). From the atlas it can be seen that as this lower level temperature wave develops, advection of cold and warm air masses by the wave amplifies the trough and ridge of the upper level geopotential wave. The resulting pressure gradients force ascent and descent in the warm and cold sectors respectively. At later stages of the eddies life cycle, as the local shears of wind and temperature are altered by the wave, the upper and lower level low pressure regions become aligned and the positive feedback is lost, resulting in barotropic decay in wave amplitude [Simmons and Hoskins, 1978].

These cyclones may also be viewed from the perspective of movements of air masses in the system: the warm conveyor belt, cold conveyor belt, and dry intrusion [Browning and Roberts, 1994]. Figure 1.8 illustrates this conceptual model, showing the movements of the air masses relative to the pressure minimum. The warm conveyor belt is a poleward moving mass of warm air, which ascends over colder air in the cold conveyor belt at the warm front, and is associated with latent heat release. The cold conveyor belt remains at low levels, and moves cyclonically around the low pressure centre. Descent from upper levels behind the cold front results in dry air moving down to lower levels in the cold sector, referred to as the dry intrusion. These motions can

\(^1\)see also p230, Holton [2004]
be identified in Figure 1.7, with ascent and descent at the warm and cold fronts respectively, and cyclonic flow at low levels. Current research suggests that latent heat release may enhance cross isentropic flow in the warm conveyor belt, and additionally strengthen the cold conveyor belt due to the PV anomaly induced by the diabatic heating [Schemm et al., 2013, Schemm and Wernli, 2014].

1.4.2 ENERGETICS

Lorenz [1955] considered the energetics of the atmosphere, partitioning this between available potential and kinetic energy. The dry available potential energy is the potential energy that could be released if all the air parcels in the atmosphere were rearranged so that density and pressure contours coincide. This gives an estimate for the energy available for conversion to kinetic energy via the rearrangement of air parcels when density surfaces are not horizontal (i.e. fluid is baroclinically unstable) or if the fluid is statically unstable, but it neglects the effects of latent heat release.
Figure 1.9: Eddy kinetic energy vs mean available potential energy for simulations with a moist model varying longwave optical thickness (circles) and insolation gradient (asterisks). Taken from O’Gorman and Schneider [2008b] (Figure 5). ©American Meteorological Society. Used with permission.

Available potential and kinetic energies can be further divided into energy associated with the mean state and with eddies. The conversion of one type of energy to the other can then be used as a different way to picture the atmospheric circulation. In this framework, the energy imbalance between low and high latitudes pictured in Figure 1.2 generates zonal available potential energy. This is converted into eddy available potential energy by the eddies, which is then either depleted by diabatic processes, or converted to eddy kinetic energy. Finally this is transformed to zonal kinetic energy via barotropic processes (barotropic decay of the wave).

Schneider and Walker [2008] and O’Gorman and Schneider [2008b] looked at the relationship between eddy kinetic energy, EKE, and zonal mean available potential energy, MAPE, in a range of climates. Based on Lorenz [1955] they calculate EKE and MAPE as:
\[
EKE = \int_{\sigma_t^{max}}^{\sigma_s} d\sigma \frac{\langle u'^2 + v'^2 \rangle}{2} \quad (1.46)
\]
\[
MAPE = \frac{c_p \rho_0}{2g} \int_{\sigma_t^{max}}^{\sigma_s} d\sigma \Gamma \left( \frac{\langle \bar{p} \rangle}{\rho_0} \right)^\kappa \left( \langle \bar{\theta} \rangle - \langle \bar{\theta} \rangle^2 \right) \quad (1.47)
\]
\[
\Gamma = -\frac{\kappa}{\langle \bar{p} \rangle} \left( \frac{\partial \bar{\theta}}{\partial p} \right)^{-1} \quad (1.48)
\]

In the above, following the notation used by O’Gorman and Schneider [2008b], \( \sigma_s \) and \( \sigma_t^{max} \) are \( \sigma = p/p_s \) at the surface and lowest tropopause levels, an overline indicates a zonal and temporal mean, and <..> is the (area weighted) mean over a baroclinic zone centred on the latitude of the maximum of the vertical integral of \( \nu \theta' \cos \phi \). These studies [Schneider and Walker, 2008, O’Gorman and Schneider, 2008b] found a linear scaling between MAPE and EKE, implying that the energy associated with eddies can be determined simply from the meridional temperature gradient in the storm track and the static stability. Figure 1.9, taken from O’Gorman and Schneider [2008b] shows this relation for a range of simulations using a moist aquaplanet. The scaling appears to hold in this model as well as in dry models, suggesting that latent heat release does not significantly alter eddy growth.

From the results of O’Gorman and Schneider [2008b], it would seem that the dominant effect of latent heat release on EKE will be due to changes to the zonal mean thermal structure, rather than due to heat release in the wave. However, results of baroclinic lifecycle experiments, in which waves develop on a prescribed initial flow, suggest a strengthening of storms as moisture is increased [Gutowski Jr. et al., 1992, Booth et al., 2013]. For given initial conditions latent heat release has been demonstrated to strengthen storms’ circulations but shorten their lifecycles. Increases in the magnitudes of EKE and the storms’ central pressure minima have been observed,
although this effect is relatively small under a similar increase in humidity to that seen under global warming simulations [Booth et al., 2013]. From these experiments it seems possible that latent heat release may change the rate of conversion of MAPE to EKE, in addition to the amount of MAPE available for conversion.

1.4.3 Eddy heat fluxes

In addition to energising storms, increased moisture may affect the transport of heat, changing the balance and the mechanisms involved [Trenberth and Stepaniak, 2003]. In particular it is clear that it can be expected that more energy be carried in the form of latent heat. Eddy heat transport by the atmosphere dominates the midlatitude global energy flux [Trenberth and Caron, 2001], and has been demonstrated to be sensitive to climate [O’Gorman and Schneider, 2008a]. The moist static energy, $H$, is defined:

$$H = c_p T + gz + Lq$$

(1.49)

where the first two terms account for the dry static energy of an air parcel, and the final term accounts for the latent heat carried by the water vapour [Neelin and Held, 1987]. The eddy flux of moist static energy is then $\nu' H'$. In a model with fixed optical depths and a mixed layer aquaplanet boundary condition, Frierson et al. [2007a] found that as moisture was increased, the resulting increase in latent energy transport was compensated by a decrease in dry static energy transport. A diffusive energy balance model was used to interpret the importance of static stability, diffusivity coefficient, and emission level to this result, but the study did not examine the systems in which the heat is carried.

The warm conveyor belt is generally considered to be responsible for the majority of heat transport in a cyclone [Eckhardt et al., 2004, Madonna et al., 2014]. When only
Figure 1.10: Composite pressure vs longitude colour map of meridional heat transport, W/1000hPa, for extreme $v' H'$ events. Diagonal striping indicates equatorward heat transport. Cross hatching indicates regions that are not statistically significant at the 99% confidence level. Continuous contours show meridional velocity anomalies (spacing 5 m/s), dashed contours moist static energy anomalies (spacing 3 K). Thicker contours correspond to negative values. (a) and (b) show events corresponding to positive and negative meridional velocity anomalies at the location of the $v' H'$ maximum respectively. The data cover latitudes north of 30°N, and includes DJFs from December 1989 to February 2011. Taken from Messori and Czaja [2015] (Figure 3). ©2015 Royal Meteorological Society. Used with permission.

Condensation and evaporation are considered as sources of diabatic heating, the warm conveyor belt has been shown to strengthen in an idealised model [Schermm et al., 2013]. In an idealized baroclinic-wave simulation, Boutle et al. [2011] found that increasing the absolute temperature of the background state, and so altering the saturation vapour pressure, increased both storm EKE and moisture ventilation by the warm conveyor belt and shallow convection, implying increased latent energy transport.

In many heat transport studies, much of the focus has been on the warm sector of the storm and its amplification when latent heating is included, with the assumption
that this sector is the key to understanding the transports by the system. However, in a study of midlatitude heat transport by extreme events in the ERA-Interim dataset, Messori and Czaja [2015] found that approximately 40% of these events are associated with equatorward movement of cooler air. Figure 1.10 shows their composites of extreme $v'H'$ events, defined as values of $v'H'$ exceeding the 95th percentile of the distribution, associated with positive (a) and negative (b) values of $v'$. Despite the difference in sign of the velocity and heat anomalies, the structures are similar. They suggest that in the positive anomaly case, the warm conveyor belt may be responsible for the extreme heat transport, although this may not be true for every extreme. In Chapter 5, changes to heat transport by both the warm and cold sectors of the storm will be investigated by varying the moisture content of the MITgcm.
2

Set-up and experiments

This chapter will give an overview of the technical details of the set-up used and experiments performed. Firstly, a summary of the MITgcm will be provided. Following this, the experiments that will be discussed in Chapters 4 and 5 will be outlined. An example of the default equilibrium climate will then be given to provide context for the results chapters. Chapter 3 of the thesis details the parameterisation of a radiation scheme for use in simple climate models. This scheme is fitted by comparison with a fast radiative transfer code, SBDART, which will be described in Section 2.4 of this chapter, followed by an overview of existing simple parameterisations.
2.1 THE MITgcm

The MITgcm is an intermediate complexity climate model. The dynamical equations are solved in full, following Marshall et al. [1997a,b], but physical processes; radiative transfer, friction, latent heat release, are approximated by relatively simple parameterisations. This framework allows idealised studies to be performed, where boundary conditions (e.g. insolation, SST) or the parameterisation of these physical processes are varied. The effects of various diabatic processes on the circulation can therefore be isolated and investigated.

2.1.1 DYNAMICAL CORE

For the experiments presented in this thesis, the MITgcm is run as a hydrostatic aquaplanet atmosphere-only model on 25 equally spaced pressure levels. The primitive equations solved by the model are Equations 1.1 to 1.7 (see previous chapter). At the upper boundary a no normal flow condition is imposed, so that here $\omega = 0$, while the lower boundary is a moving nonlinear free surface [Campin et al., 2004].

To aid in atmosphere-ocean coupling, the MITgcm takes advantage of the mathematical isomorphisms governing the evolution of the two systems [Marshall et al., 2004]. The governing equations are solved in a generic vertical coordinate, r, corresponding to rescaled pressure and height in the atmosphere and ocean respectively [Adcroft and Campin, 2004]. In the atmosphere, the case relevant here, rescaled pressure is defined as $p^* = (p/p_s)p_0$, where $p_s$ is the instantaneous surface pressure and $p_0$ a reference surface pressure (1000 hPa). Data is then output in pressure coordinates.

The equations are solved on a cube-sphere with resolution C32, i.e. 32 x 32 points on each face. This corresponds to a resolution of roughly 2.8\degree at the equator [Adcroft et al., 2004]. The cube-sphere gridding has the advantage over spherical coordinates of
avoiding having the equations solved converging at the same location as the grid (i.e. at the poles). The more uniform grid box sizes allow higher resolutions to be used without requiring as short a time step as the latitude-longitude grid would demand. However, in place of two poles, there are now four corners in the midlatitudes of each hemisphere. Taking wavenumber spectra confirms that these do not affect the storm track, with harmonics of wavenumber four not clearly dominant over other wavenumbers (e.g. Figure 4.22, Chapter 4).

As outlined in the previous chapter, the use of the cube-sphere grid requires that vector-invariant forms of the momentum equations be used, for example Equation 1.38, so that the equations solved are not dependent on the gridding. For plotting, the equations are regredded in latitude-longitude coordinates, with vectors rotated to align with spherical coordinates. These procedures are performed by MATLAB functions included as part of the MITgcm package (cube2latlon.m and rotate_uv2uvEN.m*). For accuracy, rather than evaluate a more ‘classic’ momentum budget using the regredded variables [e.g. Haigh et al., 2005], it is better to use the regredded terms of the vector-invariant momentum budget (Equation 1.38). The closure achieved will be presented with the model climatology later in this chapter.

The use of pressure coordinates means that every model layer has the same thickness, simplifying calculation of vertically integrated quantities. The use of a nonlinear free surface ensures that tracers are conserved [Campin et al., 2004]. A reference geopotential is defined as:

\[ \Phi_0 = \int_p^{p_0} \frac{1}{\rho_0} dp \]

\[ \rho_0 = \frac{p}{RT_0} \]

*Available at http://mitgcm.org/viewvc/MITgcm/MITgcm/utils/matlab/cs_grid/
where \( T_0 \) is a reference temperature profile input to the model at initialisation (see Section 2.2, below). The geopotential anomaly relative to this reference state is calculated by the model as two components: a surface contribution relative to a reference pressure (1000 hPa), and a hydrostatic anomaly relative to the surface pressure (see MITgcm manual, Adcroft et al. [2015]):

\[
\Phi' = \int_{p_0}^{p_{surf}} \frac{1}{\rho_0} dp + \int_p^{p_{surf}} \left( \frac{1}{\rho} - \frac{1}{\rho_0} \right) dp
\]

For comparison with other data sets, for example reanalysis data, the total geopotential relative to the reference pressure is useful. This can be evaluated by combining the reference geopotential, hydrostatic anomaly, and surface terms output by the model.

### 2.1.2 Physics parameterisations

The physics module used follows O’Gorman and Schneider [2008a] and is based on Frierson et al. [2006] and Frierson [2007]. This includes parameterisations of radiative transfer, surface fluxes of stress, sensible heat, and moisture, diffusive processes, large-scale condensation, and moist convection.

Vertical diffusion occurs within the model boundary layer, the depth, \( h \), of which is determined by the value of a bulk Richardson number:

\[
Ri(z) = \frac{g z [\theta_v(z) - \theta_v(z_a)]/\theta_v(z_a)}{[v(z)]^2}
\]

where \( z_a \) is the height of the lowest model level. The boundary layer is defined to be where this Richardson number has a value of less than \( Ri_c = 1 \), indicating that turbulent processes are of more importance than static stability. The diffusion and drag coefficients for the surface fluxes and boundary layer diffusion are calculated using the scheme described in O’Gorman and Schneider [2008a], based on Troen and
Mahrt [1986], with a roughness length of 0.05 m for momentum, moisture and sensible heat. Diffusion is both of deeper extent and stronger where static stability is lower.

The model’s diffusive heating term can be expressed:

\[ Q_{\text{diff}} = \frac{\partial}{\partial z} \left( K \frac{\partial \theta}{\partial z} \right) \]  \hspace{1cm} (2.5)

The diffusive heating is therefore related to both the curvature of the potential temperature profile and, through \( K \), to the local stability of the atmosphere. Warm air will tend to be cooled and cold air warmed.

The model uses an approximate Clausius-Clapeyron equation with a constant latent heat of vaporisation (cf. Equation 1.27):

\[ e_s = e_{s0} e^{-(L/R_v)((1/T) - (1/T_0))} \]  \hspace{1cm} (2.6)

In the above, \( e_{s0} \) is a reference value of saturation vapour pressure at a reference temperature \( T_0 = 273.16 \) K. \( L \), the latent heat of vaporisation of water, is taken to have a constant value of \( 2.5 \times 10^6 \) Jkg\(^{-1}\).

Large-scale condensation occurs when a grid box becomes saturated, i.e. has a vapour pressure above the saturation vapour pressure. Clouds are not included, with condensate instead falling straight out with no re-evaporation. Upon saturation, the specific humidity is adjusted as (see Equation 21 of Frierson et al. [2006]):

\[ \delta q = \frac{q_s - q}{1 + \frac{L a_{qs}}{c_p a_q}} \]  \hspace{1cm} (2.7)

where \( q_s \) is the saturation specific humidity. Moist convection is parameterised using a simple Betts-Miller scheme [Frierson, 2007, O’Gorman and Schneider, 2008a], which will be summarised here. For further detail see Frierson [2007]. Firstly, temperature
and humidity are relaxed towards reference profiles:

\[ \delta q = -\frac{q - q_{ref}}{\tau_{SBM}} \]  \hspace{1cm} (2.8)

\[ \delta T = -\frac{T - T_{ref}}{\tau_{SBM}} \]  \hspace{1cm} (2.9)

Temperature is relaxed to a dry adiabat up to the lifting condensation level (LCL), and a moist adiabat above this. Humidity is relaxed towards a relative humidity of 70% relative to this reference temperature. In both cases the relaxation time, \( \tau_{SBM} \), is 2 hours.

The height of convection is determined by integrating the above from the surface to the level of zero buoyancy (LZB), allowing precipitation due to drying and warming to be calculated:

\[ P_q = -\int_{p_0}^{p_{LZB}} \delta q dp / g \]  \hspace{1cm} (2.10)

\[ P_T = \int_{p_0}^{p_{LZB}} \frac{c_p}{L} \delta T dp / g \]  \hspace{1cm} (2.11)

Where the \( P_q \) is positive, the column contains more moisture than the reference profile. Where \( P_T \) is positive, the convectively available potential energy is greater than the convective inhibition.

For the case where both \( P_T \) and \( P_q \) are positive, the temperature and humidity profiles are relaxed up to the LZB. Following O’Gorman and Schneider [2008a], where the adjustment of the humidity profile results in a greater enthalpy change than that of the temperature profile, enthalpy conservation is achieved by lengthening the time scale of the humidity adjustment.
Where \( P_T \) is positive, but \( P_q \) is not, a pressure level, \( P_{shall} \), is found such that:

\[
P_q = \int_{P_0}^{P_{shall}} \left( \frac{q - q_{ref}}{\tau_{SBM}} \right) dp = 0
\]  

(2.12)

The temperature profile is then relaxed up to this height.

The radiation scheme has not previously been a key area of focus in the development of this simple physics package. In the existing module, longwave radiative transfer can be included following Frierson et al. [2006], or with a parameterisation of optical depth dependent on specific humidity taken from Byrne and O’Gorman [2013]. Shortwave radiation does not interact with the atmosphere. These parameterisations are detailed in Section 2.4.2, below. In Chapter 3 an alternative scheme is produced, used subsequently in Chapters 4 and 5.

The SSTs are zonally invariant and fixed with a latitude distribution (°C) [Williamson and Coauthors, 2012]:

\[
T_s(\phi) = 27 \left[ 1 - \sin^2 \left( \frac{90}{60} \phi \right) \right] \quad |\phi| < 60°
\]

(2.13)

\[
T_s(\phi) = 0 \quad |\phi| \geq 60°
\]

(2.14)

The top of atmosphere insolation used is a perpetual equinox with no seasonal or diurnal cycle [O’Gorman and Schneider, 2008a]:

\[
R_s = R_{s0} [1 + \Delta_s p_2(\phi)]
\]

(2.15)

\[
p_2(\phi) = \frac{1}{4} [1 - 3 \sin^2(\phi)]
\]

(2.16)

where \( R_s \) is the insolation, \( R_{s0} = 1360.0/4 \text{ Wm}^{-2} \), and \( \Delta_s = 1.4 \). Each model month is taken to be 30 days long, so that a year refers to a period of 360 days. With no seasonal
## Table 2.1: Summary of experiments performed with the MITgcm, including symbols used for plotting data

<table>
<thead>
<tr>
<th>Reference name</th>
<th>$e_{50}$ fraction</th>
<th>Description</th>
<th>Plotting symbol</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>rad-on</em></td>
<td>1.0</td>
<td>Optical depths calculated from humidity</td>
<td>○</td>
</tr>
<tr>
<td><em>dry</em></td>
<td>0.0</td>
<td></td>
<td>+</td>
</tr>
<tr>
<td><em>010</em></td>
<td>0.1</td>
<td></td>
<td>*</td>
</tr>
<tr>
<td><em>025</em></td>
<td>0.25</td>
<td>Fixed optical depths, varying humidity</td>
<td>▽</td>
</tr>
<tr>
<td><em>050</em></td>
<td>0.5</td>
<td></td>
<td>×</td>
</tr>
<tr>
<td><em>075</em></td>
<td>0.75</td>
<td></td>
<td>△</td>
</tr>
<tr>
<td><em>wet</em></td>
<td>1.0</td>
<td>Fixed optical depths, SST perturbed in the tropics</td>
<td>☆</td>
</tr>
<tr>
<td><em>125</em></td>
<td>1.25</td>
<td></td>
<td></td>
</tr>
<tr>
<td><em>150</em></td>
<td>1.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td><em>tropheat</em></td>
<td>0.5</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

cycle and the aquaplanet set-up, all days and both hemispheres are equivalent.

### 2.2 Experiments

Ten eleven-year simulations have been performed, summarised in Table 2.1. The atmosphere is stationary when initialised, with zonally symmetric prescribed potential temperature and humidity fields, shown in Figure 2.1. The fixed SST means that the model equilibrates quickly. Figure 2.2 shows the mean kinetic energy of the model, averaged over all latitudes and pressure levels, as a function of time for the first three years of the *wet* experiment. Even in steady state, the scatter is large, with a standard deviation of 2.5 kJm$^{-2}$. Initially there is a bias towards large values of kinetic energy, sometimes over 4.5 standard deviations away from the temporal mean, as the model responds to the initial temperature field. By approximately day 100 the data appears to be distributed symmetrically around the mean, indicating that an equilibrium state has been reached. Although the variability is still large, after the spin-up period all data lies within 4 standard deviations of the mean. The first year of data

\[\text{For the dry run, the initial humidity profile is set to zero, so that the atmosphere is completely dry.}\]
Figure 2.1: (left) Potential temperature, K, and (right) specific humidity, g/kg, profiles used for model initialisation.

Figure 2.2: Pressure and area averaged kinetic energy for the first 3 years of the wet experiment. Solid line indicates the time mean over the 10 years used for analysis.
for each experiment is therefore discarded as spin-up, and the remaining ten years are analysed.

The *rad-on* experiment uses the radiation scheme presented in Chapter 3 to calculate optical depths from the specific humidity. This allows both a test of the performance of the scheme, and calculation of optical depths for use in the remaining experiments. Temporal and zonal averages of the optical depths calculated in this run are taken and are used as a fixed input for the remaining experiments. This allows the effects of latent heat on the model dynamics to be explored separately from the radiative effects, but with a more useful control climate than that found with the existing parameterisations in the model.

Humidity is varied by multiplying the constant $e_{s0}$ in Equation 2.6 by factors of 0, 0.1, 0.25, 0.5, 0.75, 1, 1.25, and 1.5 [Frierson et al., 2006]. This means that for a given temperature the amount of moisture held by the atmosphere, and the resulting latent heat release when this condenses, is scaled. For the *dry* run moist parameterisations are switched off to ensure that only the dry heating processes are called. The resulting time mean climates are presented in Chapter 4. Consistent behaviour between the
Table 2.2: Latitude of peak $\bar{v}/\bar{T}'$ to be used to indicate storm-track position, and tropopause and boundary layer height pressure levels at this latitude.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Peak $\bar{v}/\bar{T}'$ lat (°)</th>
<th>Tropopause pressure (hPa)</th>
<th>Boundary layer top (hPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>dry</td>
<td>39</td>
<td>220</td>
<td>740</td>
</tr>
<tr>
<td>010</td>
<td>39</td>
<td>220</td>
<td>780</td>
</tr>
<tr>
<td>025</td>
<td>39</td>
<td>220</td>
<td>820</td>
</tr>
<tr>
<td>050</td>
<td>41</td>
<td>220</td>
<td>860</td>
</tr>
<tr>
<td>075</td>
<td>41</td>
<td>180</td>
<td>860</td>
</tr>
<tr>
<td>wet</td>
<td>43</td>
<td>220</td>
<td>900</td>
</tr>
<tr>
<td>125</td>
<td>47</td>
<td>220</td>
<td>900</td>
</tr>
<tr>
<td>150</td>
<td>45</td>
<td>220</td>
<td>900</td>
</tr>
</tbody>
</table>

experiments is seen, including the dry run, justifying this set-up.

Varying the model’s moisture content results in very different mean climates between experiments. For diagnostics relating to midlatitude eddies, the latitude of maximum $\bar{v}/\bar{T}'$ will therefore be used as an indicator of storm-track position, rather than assuming this to be fixed. Also useful for intercomparison of the dynamics of these very different models are the tropopause and boundary layer height levels. These values for each run are shown in Table 2.2 for the storm track latitude. For tropopause height the WMO definition is used, so that this is taken to be the lowest pressure level where the zonal mean lapse rate reaches, and stays below, 2Kkm$^{-1}$ [WMO, 1957]. The boundary layer height is approximated as the zonal mean of that used in the diffusion scheme, described above, taken to the nearest pressure level.

The final experiment, tropheat, includes a perturbation to the SST in the tropics:

$$T_s'(\phi) = 5 \cos^2 \left( \frac{90}{10} \phi \right) \quad \text{for } |\phi| < 10^\circ \quad (2.17)$$

$$T_s'(\phi) = 0 \quad \text{for } |\phi| \geq 10^\circ \quad (2.18)$$

The control and perturbed SSTs are shown in Figure 2.3. One effect of increasing humidity is that the already warm tropics moisten, and are then warmed further by la-
tent heat release (see Chapter 4). Increasing the tropical SST approximately reproduces this effect without varying \( e_{a0} \), leading to an increase in the equator-pole temperature difference, but a less altered static stability, particularly in the midlatitudes. This set-up allows some investigation of how much of the change to the circulation is driven from the tropics alone.

### 2.3 Climatology

To give an example of the climate observed in the experiments, Figure 2.4 shows a range of zonal mean fields from the *wet* simulation. This is the default climate for the experiments in Chapters 4 and 5 of this thesis, with fixed optical depths and a reference saturation vapour pressure of 610.78 Pa, which is the default value for the model and most comparable to the real world. To allow comparison with Figure 1.1, the same fields have been plotted: zonal wind speed, temperature, meridional wind speed, vertical wind speed, specific humidity, and relative humidity.

Two zonal jets are seen; a subtropical jet, balancing advection of vorticity by the meridional wind, and an extratropical, eddy driven jet. The structure of the Hadley cell can be seen from the meridional and zonal winds, with ascent in the tropics, poleward movement of air aloft, descent in the subtropics, and a return flow along the surface. In the midlatitudes, equatorward motion is seen at upper levels, while surface winds are poleward, indicating the presence of an eddy-driven Ferrel cell. The meridional streamfunction will be shown in Chapter 4, wind speeds are shown here to allow comparison with Figure 1.1.

Looking at the specific humidity, it can be seen that the moistest air is found in the tropics, associated with warmer temperatures, as indicated by Equation 1.27. The advection of water vapour by the meridional circulation is reflected by the plume of high specific humidity in the tropics, carried upwards by the ascending branch of the
Figure 2.4: Zonal mean climatology of wet experiment. Top row: zonal wind speed, m/s, temperature, K, and meridional wind speed, m/s. Bottom row: vertical wind speed, hPa/day, specific humidity, g/kg, and relative humidity. Data is averaged over 10 years and over both hemispheres.
Figure 2.5: Vertical integral of the terms in the angular momentum budget (Equation 1.40) for each run over all model levels. Solid line represents $C_{total}$, dot-dashed line $C_{zonat}$, dashed line $C_{eddy}$, cyan line (-1×) surface stress, and red line budget residual. Horizontal dashed line indicates the global average of the surface wind stress.
Hadley cell. Looking at the relative humidity, it can be seen that the tropics are near saturation up to the tropopause, above which the air is very dry. By contrast, the subtropics are relatively dry, with much of the water vapour having been precipitated out in the saturated tropics. The air warms adiabatically as it descends, increasing the saturation vapour pressure and acting to decrease the relative humidity. The poles, where temperature, and consequently saturation vapour pressure, is low, are again closer to saturation. Close to the surface, in the boundary layer, where turbulent processes diffuse water vapour upwards from the sea surface, the atmosphere is also near-saturated.

Comparing Figure 2.4 with Figure 1.1, it can be seen that many features of the climatology are similar to the multi-model mean of the APE. Key differences are the double jet structure seen in this experiment, and the wider Hadley cell. However, the temperature and humidity structures are generally similar. The processes driving the structures of the different variables shown here will be discussed in Chapter 4.

Figure 2.5 shows the zonal mean momentum budget, vertically integrated over all pressure levels for each experiment (see Equation 1.40). In equilibrium, the total momentum convergence, $C_{total}$, is expected to balance the surface stress, $\tau_S$, at each latitude. Comparing the solid black and cyan lines shows that the calculation of the momentum budget captures this balance well, with only a small residual (red line). As the budget closes perfectly when terms are added in cube-sphere coordinates, this residual is attributed to errors associated with rotation of vectors and regridding to spherical coordinates. Linearising about the zonal mean state, $C_{total}$ can be divided into the contribution of eddies, $C_{eddy}$ (dashed line), and that of the mean flow, $C_{zonal}$ (dot-dashed line). It can be seen that $C_{eddy}$ is the dominant contribution to $C_{total}$ at most latitudes, emphasising the importance of eddies in the atmosphere.

In equilibrium, it is expected that the net transfer of angular momentum into the
atmosphere should be zero. The horizontal line in Figure 2.5 indicates the global average of the surface wind stress. In the drier runs this lies close to zero as expected. As moisture increases, the magnitude of the mean wind stress increases. This implies that the momentum of the atmosphere should be increasing with time, suggesting that a steady state has yet to be reached.† However, when the global mean surface wind stress is plotted as a function of time for the full wet run (not shown), after the spin-up period it remains steady around a value of approximately 0.035 Nm⁻², as indicated by Figure 2.5. Further, it was shown in Figure 2.2 that the total kinetic energy of the atmosphere remains relatively steady after an initial year of spin-up. A second possibility is that there is a missing contribution to the surface torque budget, related in some way to the moisture content of the model, that needs to be considered. Egger et al. [2007] summarise the terms in the angular momentum budget in detail, including the contributions of the moisture and orography torques. The former results from the transfer of mass polewards due to the dominance of evaporation (precipitation) at lower (higher) latitudes, and can be expressed:

\[ T_q = \int (E - P)\Omega r^4 \cos^3 \phi d\lambda d\phi \tag{2.19} \]

As would be expected, the contribution of this term to the budget of torques on the atmosphere increases as moisture increases. This contribution is, however, still small compared with the imbalance due to the frictional torque, and does not explain the apparent imbalance. The orography torque due to deformation of the nonlinear free surface is also negligible. Further investigation is underway into possible sources for the imbalance, for example potential loss of angular momentum due to noise filtering. As the atmosphere does appear to reach a steady state, it is hoped that this issue does

†Note that it is the negative of the surface stress is plotted, so that negative values imply an acceleration.
not strongly affect the conclusions of the thesis.

2.4 Radiation modelling

In Chapter 3, simple parameterisations of long and shortwave radiative transfer are tested and improved upon by comparison with the Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) program. A quick summary of the code, and of the existing radiation schemes tested, are provided here.

2.4.1 SBDART

SBDART is a fast code for radiative transfer in the Earth’s atmosphere, described in detail by Ricchiazzi et al. [1998]. The molecular absorption component uses the band models developed for LOWTRAN 7 [Pierluissi and Peng, 1985], and has a spectral resolution of 20 cm$^{-1}$. Rayleigh scattering coefficients are calculated as a function of wavelength [Shettle et al., 1980, Liou, 2002]. The program additionally includes parameterisations of Mie scattering, radiative effects of aerosols and ground reflectance properties. However, as the MITgcm does not include parameterisations of cloud or aerosol, and uses a fixed surface albedo, these are not used in testing its radiation scheme. The radiative transfer equation is solved approximating plane-parallel radiative transfer. Ricchiazzi et al. [1998] compared SBDART calculations of radiance with observations from the Spectral Radiation Experiment (SPECTRE) [Ellingson and Wiscombe, 1996] and from the Baseline Surface Radiation Network (BSRN). They found that the code performed well across the spectrum, with very good agreement in the longwave, and the shortwave also good, although affected by uncertainties in the treatment of aerosols.

SBDART can take a user-prescribed atmospheric profile as an input. To obtain results useful for the conditions observed in the MITgcm, temperature, humidity, pres-
sure and altitude profiles were derived from model output and the MITgcm fixed SST and albedo were used as the surface condition. Cloud, aerosols and ozone were not accounted for as the intended purpose of the output is solely to parameterise absorption by water vapour. Default SBDART fractions of well mixed trace gases were, however, used. The default inputs for SBDART can be found in Ricchiazzi [2002]. To aid reproducability, the non-default settings used here are summarised in Table 2.3. The code was run for four spectral ranges: shortwave (0.25 - 4.0\(\mu m\)), longwave (4.005 - 7.995 and 14.005 - 100.0\(\mu m\)) and window (8.0 - 14.0\(\mu m\)). For the settings used, for a given input profile, SBDART calculates spectrally integrated downwelling and upwelling fluxes and heating rates, which can be compared with those derived from the parameterisations used in the MITgcm.

### Table 2.3: Summary and explanation of non-default values of input parameters used in the SBDART runs. See Ricchiazzi [2002] for full list of parameters and defaults.

<table>
<thead>
<tr>
<th>Input name</th>
<th>Input used</th>
<th>Explanation</th>
</tr>
</thead>
<tbody>
<tr>
<td>wlinf</td>
<td>(see text)</td>
<td>Lower wavelength limit (set to minimum allowed)</td>
</tr>
<tr>
<td>wlsup</td>
<td>(see text)</td>
<td>Upper wavelength limit (set to maximum allowed)</td>
</tr>
<tr>
<td>wlinf</td>
<td>0.005(\mu m)</td>
<td>Spectral resolution of SBDART run</td>
</tr>
<tr>
<td>idatm</td>
<td>0</td>
<td>Use user supplied atmospheric profile</td>
</tr>
<tr>
<td>iout</td>
<td>11</td>
<td>Output fluxes integrated over wavelength</td>
</tr>
<tr>
<td>albcon</td>
<td>0.06</td>
<td>Spectrally uniform surface albedo value</td>
</tr>
<tr>
<td>btemp</td>
<td>Model SST (Equation 2.13)</td>
<td>Input sea surface temperature, K</td>
</tr>
</tbody>
</table>

2.4.2 Simple radiation schemes

Two longwave radiation parameterisations have currently been incorporated into the MITgcm grey atmosphere package. The first follows Frierson et al. [2006], and solves the two stream approximation using fixed optical depths. These are prescribed to depend simply on pressure and latitude, and are calculated as:
\[
\tau_0 = \tau_{0e} + (\tau_{0p} - \tau_{0e})\sin^2(\theta)
\]
(2.20)

\[
\tau = \tau_0 \left[ f_l \left( \frac{p}{p_s} \right) (1 - f_l) \left( \frac{p}{p_s} \right)^4 \right]
\]
(2.21)

where \( \tau \) is optical depth, equivalent to \( \int kpds \) in Equation 1.26, \( \tau_{0e} = 6.0 \), \( \tau_{0p} = 1.5 \), and \( f_l = 0.1 \).

The second parameterisation is that used by Byrne and O’Gorman [2013] in a study on land-ocean warming contrasts. This uses a much simplified version of the Beer-Lambert equation, with optical depths linearly dependent on water vapour over the whole longwave spectrum:

\[
\frac{d\tau}{d\sigma} = a\mu + bq
\]
(2.22)

Here \( a \) and \( b \) are nondimensional constants with values 0.8678 and 1997.9 respectively. \( \mu \) represents the abundance of non-water vapour greenhouse gases and takes a default value of 1. However, these parameterisation coefficients were chosen for ease of comparison with the prescribed optical depths previously used in their model, and as a simple way of including radiative feedbacks. They can therefore not be expected necessarily to provide a realistic approximation of the radiative effects of water vapour.

Absorption of shortwave radiation by the atmosphere has not yet been incorporated into the MITgcm grey atmosphere package. An approximation to a seasonal and diurnal average insolation (Equation 2.15) is however included, useful where a slab ocean is coupled to the atmosphere.

In Chapter 3 the optical depths from these simple longwave parameterisations will
be compared against those calculated using SBDART. Additionally, test runs of the model are performed with these longwave schemes, which are shown to not provide an ideal circulation structure. The remainder of Chapter 3 therefore discusses the development of an improved simple parameterisation of radiative processes for the MITgcm.
Parameterisation of the Radiative Effects of Water Vapour

In our initial set-up, the MITgcm did not include any representation of atmospheric absorption of shortwave radiation by water vapour, while the longwave schemes included had not been compared with other radiative transfer codes. To provide a more realistic description of the radiative effects of water vapour in the model, in this chapter the current parameterisations will first be compared with output from a more comprehensive radiative transfer code, SBDART (see Section 2.4.1). The heating rates calculated using these schemes will be shown to provide a poor approximation to water vapour radiative heating. A new simple radiation scheme will then be presented, constructed by fitting to SBDART optical depth profiles for given temperatures and humidities.
Finally, the effects of incorporating this scheme into the MITgcm will be discussed.

3.1 SBDART results and testing of longwave models

As described in Section 2.4.2, two parameterisations of longwave radiation are currently incorporated into the MITgcm. In order to test the validity of these models, we compare the optical depths and heating rates they generate with those of SBDART, a fast radiative transfer code (see Section 2.4.1). For the intercomparison of the different radiation schemes to be discussed in this chapter, the specific humidities and temperatures shown in Figure 3.1 will be used. These are zonal means from the rad-on run of the MITgcm, averaged over 10 years. The peak temperatures and specific humidities are colocated in the tropics, as would be expected from the Clausius-Clapeyron equation. Specific humidity decreases quickly with latitude, with little moisture poleward of 50°. Temperature decreases with height, with minima at approximately 100hPa over the equator and 200hPa over the poles. Above this, in the model stratosphere, temperature increases with height, with small peaks at 50° latitude. The model physics package excludes clouds, with water precipitated out immediately following condensation. Consequently, only clear sky conditions are parameterised here, although ideas to include cloud will be discussed in Section 6.2.2.

In this chapter, in accordance with the SBDART runs described in Section 2.4.1, shortwave will refer to wavelengths of 0.25 to 4\mu m, and longwave to wavelengths from 4 to 100\mu m. In the shortwave, the downwelling fluxes output from SBDART for a default insolation of 1363.3kWm\(^2\) are used to evaluate transmission coefficients for the input humidity profile. These are then used with the MITgcm perpetual equinox insolation, Equation 2.15, in order to evaluate heating rates comparable to those expected in the model.

Figure 3.2 shows the longwave and shortwave heating rates evaluated by SBDART
for the profiles in Figure 3.1. In the longwave (left), SBDART shows cooling throughout the atmosphere. The highest magnitude cooling occurs over the more humid lower latitudes. While the intensity of the solar spectrum is peaked in the shortwave, the blackbody spectra for the temperature range observed in the atmosphere peak in the longwave. Emission is therefore strong in this wavelength range. The greatest cooling will relate to temperature contrasts between warmer and cooler layers of the atmosphere. In the stratosphere regions of enhanced cooling associated with the warmer temperature areas at 50° latitude are found. A region of reduced cooling over the equator can also be seen, associated with a colder region around the tropopause. Where the gradient of the temperature profile increases at around 250hPa an associated strong cooling rate is seen, decreasing with pressure as the temperature gradient decreases. Strong cooling is also seen at the surface where there is a contrast between the lowest pressure level temperature and the SST. The SBDART shortwave heating rates (right) again show stronger heating near the equator where the specific humidity is higher, with a peak in the tropics at around 350hPa. More strongly absorbed wavelengths in the incident beam are quickly attenuated higher in the atmosphere, leading to this up-
per level warming. As the contribution of these wavelengths to the total beam lessens, the overall absorption rate of the beam decreases, so that less heating is observed at lower levels.

The heating rates from the current MITgcm implementations of longwave radiative transfer are shown in Figure 3.3. As described in Section 2.4.2, the Frierson et al. [2006] scheme uses optical depths as a function of latitude and pressure to approximate the distribution of water vapour (Equation 2.21). Comparing the heating rates resulting from this scheme (left panel) with those from SBDART shows major discrepancies in both the distribution and magnitude. Cooling is stronger at low latitudes, and reduced cooling (in this case, warming) is associated with the cool temperatures at upper levels centred over the equator, as observed in the SBDART output. Beyond these features, the parameterisation captures the predicted rates poorly, with a strong gradient in cooling rate throughout the troposphere, and a large peak at around 650 hPa in the tropics which is not seen in SBDART. The heating rates calculated using the parameterisation used in Byrne and O’Gorman [2013] (Equation 2.22) also perform poorly. The parameterisation was designed to reproduce the Frierson et al. [2006]
heating rates and so the structures are similar, with a strong cooling peak at mid levels in the tropics. The dependence on water vapour is reflected by the meridional structure, with strong gradients at lower latitudes.

To test the impact of these two schemes, the MITgcm has been run for 18 months with each. The resulting zonal mean jet and temperature structures averaged over the last 6 months of these runs are shown in Figure 3.4. Strong subtropical jets are observed for both models, but these do not extend to the surface, with a weak eddy-driven midlatitude jet resulting in very weak surface winds. Additionally, the tropics superrotate, with strongly positive zonal wind speeds of greater angular momentum than the surface found at 100 hPa. Figure 1.1 shows the zonal wind and temperatures of the multi-model mean of the Aquaplanet Experiment (APE) [Williamson and Coauthors, 2012]. Based on this climatology, weak surface winds and superrotation do not appear to be general features of aquaplanet climate. These may result in difficulties if the atmospheric model is coupled to an ocean. Zonal wind and temperature structure are strongly linked by thermal wind balance (Equation 1.12), and so the dynamics can be expected to be sensitive to the parameterisation of heating in the model.
Figure 3.4: Zonal mean zonal wind, m/s, (shading) and temperature, K, (contours) when the MITgcm is run with the parameterisations of (left) Frierson et al. [2006] and (right) Byrne and O’Gorman [2013].

ing the MITgcm climates with the APE mean suggests that the schemes currently in use do not capture the real structure of heating and cooling associated with a given humidity and temperature profile, and so yield some strange features in the circulation via the equilibrium temperature structures produced. An improved scheme is therefore presented and tested later in this chapter. As no parameterisation of shortwave radiative transfer is included, implementation of this will first be discussed.

3.2 Shortwave parameterisation

Shortwave radiative transfer in the atmosphere is not currently included in the MITgcm. To allow shortwave effects to be explored, a simple parameterisation has been developed for the model, presented here. An existing scheme, that of Lacis and Hansen [1974], is also outlined for comparison.

As discussed in Section 1.2.2, in the shortwave region of the spectrum, the key processes to be considered are absorption and scattering. A major approximation in the MITgcm grey atmosphere physics is to exclude clouds, simplifying the physics and allowing other processes to be investigated independently of cloud feedback. The model
uses a surface albedo of 0.06, typical for ocean [Briegleb et al., 1986]. This is small, so that there is little upwelling shortwave radiation. The effects of scattering and of absorption of upwelling radiation are thus small relative to absorption of downwelling radiation and have been neglected for simplicity.

The longwave schemes already incorporated in the MITgcm (see Section 2.4.2) were taken as a starting point. These approximate grey, two stream radiative transfer. Adapting these for the shortwave, fluxes of radiation in each layer are calculated by solving the two-stream approximation with the approximations of no absorption of upwelling radiation, and no shortwave emission:

$$dD = -Dd\tau$$

(3.1)

where $D$ is the downwelling flux of shortwave radiation (Wm$^{-2}$). This is the equivalent of Equation 1.26 for no sources of radiation and for a wider spectral range, rather than a single wavelength. The boundary conditions for the shortwave scheme are the insolation as per Equation 2.15, and an albedo of 0.06.

In order to parameterise transmission coefficients in the shortwave, SBDART was run for a range of temperature and specific humidity profiles. The resulting downwelling fluxes were inverted for each layer:

$$d\tau = -\log(D_{\text{out}}/D_{\text{in}})$$

(3.2)

resulting in profiles of a grey radiation optical depth. For consistency with the existing parameterisations (see Equation 2.22), a linear fit of $d\tau/dq$ to $q$ would be ideal. The relation between these is plotted in Figure 3.5 and is however clearly not linear. This method aims to use one band to describe the effects of many absorption lines, but the effect of the difference in absorptance between different wavelengths must also
Figure 3.5: $d\tau/d\sigma$ as a function of $q$, g/kg. Evaluated from SBDART data using Equation 3.2.

be accounted for. Strongly absorbed wavelengths will warm the atmosphere at upper levels, but not penetrate to lower levels. As these are removed from the downwelling beam, progressively more weakly absorbed wavelengths remain, leading to a reduction in absorption. The extinction coefficient, $b$, in Equation 2.22, is therefore found to not be a simple constant for this band.

In order to account for the change in the spectral composition of the incident beam as it travels through the atmosphere, the amount of material already traversed must be accounted for. A better approximation is therefore to consider $b$ to be a function of the optical depth from the top of the atmosphere to a given layer. The current formulation of the longwave scheme in the MITgcm, Equation 2.22, can be inverted for $b$:

$$b = \left( \frac{d\tau}{d\sigma} - a\mu \right) / q$$

(3.3)

To evaluate $a\mu$, the constant term describing the dependence of optical depth on well mixed gases, SBDART was run for a dry atmospheric profile. From the resulting values of $d\tau/d\sigma$, a value of 0.0596 was calculated. Substituting this into Equation 3.3
allows the extinction coefficients $b$ to be evaluated from the SBDART output. Figure 3.6 shows these as a function of $\tau$. As $\tau$ increases, $b$ decreases at a decreasing rate. A least squares algorithm* was used to fit test functions to this trend using an equatorial profile from an early run of the MITgcm. A good fit was found to be achieved by:

$$\log(b(\tau_{\sigma}^{SW})) = \frac{0.01887}{(\tau + 0.009522)} + \frac{1.603}{(\tau + 0.5194)^2}$$ (3.4)

This is plotted as the red line in Figure 3.6, which lies closest to tropical profiles similar to that used in fitting the scheme. The choice of using a tropical profile in evaluating the parameters can be justified by considering that this is the most humid region of the atmosphere, and so is most suitable for capturing water vapour radiative effects. However, for comparison, a second parameterisation of $b(\tau)$, fitted using drier extratropical profiles, is shown by the blue line:

$$\log(b(\tau_{\sigma}^{SW})) = \frac{0.00422}{(\tau - 0.00324)} + \frac{1.68}{(\tau + 0.478)^2}$$ (3.5)

The heating rates calculated using both fits will be presented in Section 3.4.

Lacis and Hansen [1974] parameterised absorption and scattering of shortwave radiation by ozone and water vapour in clear and cloudy skies. For the clear sky case of relevance here, they described water vapour absorption using the amount of precipitable water vapour in the path, $y_{wv}$, following Yamamoto [1962]:

$$A_{wv}(y_{wv}) = \frac{2.9y_{wv}}{(1 + 141.5y_{wv})^{0.635} + 5.925y_{wv}}$$ (3.6)

*MATLAB’s lsqcurvefit [Coleman and Li, 1994, 1996]
To account the dependence of absorption of temperature and pressure, the precipitable water vapour was multiplied by a function of pressure and temperature:

$$y_{wv}^{eff} = y_{wv} \left( \frac{p}{p_0} \right) \left( \frac{T_0}{T} \right)^{\frac{3}{2}}$$  \hspace{1cm} (3.7)

This parameterisation was found to be in good agreement with aircraft measurements of solar flux [Paltridge, 1973], but at the time they were unable to check the absorption coefficients any further. This scheme will also be tested in Section 3.4 for comparison with the new parameterisations above.

3.3 **Longwave Scheme**

In the longwave region of the spectrum, the dominant processes are absorption and emission. In the MITgcm, fluxes are again calculated using the two-stream approximation, this time accounting for both emission and absorption on upward and downward paths:
\[
\frac{dU}{d\tau} = (U - B) \quad (3.8)
\]
\[
\frac{dD}{d\tau} = (B - D) \quad (3.9)
\]
\[
B = \sigma_{SB}T^4 \quad (3.10)
\]

where \(U\) and \(D\) are the upwelling and downwelling fluxes of longwave radiation, \(B\) represents blackbody emission, and \(\sigma_{SB}\) is the Stefan-Boltzmann constant. The surface and top of atmosphere boundary conditions used are [as Frierson et al., 2006]:

\[
U_s = \sigma_{SB}T_s^4 \quad (3.11)
\]
\[
D_{TOA} = 0 \quad (3.12)
\]

To obtain optical depth values for the longwave from the downward SBDART fluxes, Equation 3.9 was inverted:

\[
d\tau = -\log\left(\frac{D_{out} - B}{D_{in} - B}\right) \quad (3.13)
\]

Figure 3.7 shows the resulting optical depths as a function of \(q\) for the whole longwave region. The scatter does not follow any clear functional form.

A key feature of the longwave absorptance spectrum is the infrared window, which can be seen in Figure 1.3 to span wavelengths from approximately 8 to 14 \(\mu\)m. Radiation emitted at wavelengths in this region is weakly absorbed and penetrates further through the atmosphere. The significant difference in behaviour between wavelengths
Figure 3.7: Longwave optical depth for wavelengths 4-100\(\mu m\) as a function of \(q,\) g/kg. Evaluated from SBDART data using Equation 3.13.

Figure 3.8: Black-body spectrum for the peak SST value (300.15 K). Shaded area shows the wavelengths parameterised as in the window region.
Figure 3.9: Longwave optical depth as a function of $q$, g/kg, for wavelengths (left) 4-8 and 14-100\,$\mu$m and (right) 8-14\,$\mu$m calculated from SBDART data using Equation 3.13. Bottom panels zoom in on $q$ between 0 and 2 g/kg to show fit at lower humidity. Blue lines indicate the fit achieved by Equations 3.14 and 3.15.
in the window and those over the rest of the spectrum may partially explain the complex behaviour of the longwave transmission coefficient in Figure 3.7. Figure 3.8 shows the blackbody spectrum for our peak SST, 300.15K, as a function of wavelength. The blue shaded region indicates the infrared window. Integrating over this and normalising by the total, it is found that the proportion of blackbody radiation emitted in the window region can be approximated as 37%. Figure 3.9 shows equivalent plots to Figure 3.7, but with the ‘non-window’ (left panels) and ‘window’ (right panels) optical depths evaluated separately. Optical depths are again evaluated from the downward SBDART fluxes using Equation 3.9, but with 37% of blackbody emission now assigned to the ‘window’ region. It can be seen that the window region has significantly lower absorption than the other wavelengths.

In both regions the scatter still does not show a linear relation, as a wide range of wavelengths with differing absorptance is covered. However the dependences are now simple enough to be described by functional relations. The window region can be fitted by approximating a quadratic relation, while the non-window wavelengths may be parameterised as a function of $q^{1/2}$. The resulting fits are:

\[
\frac{d\tau_{LW}}{d\sigma} = 0.15493 + 351.48q^{1/2} \quad (3.14)
\]

\[
\frac{d\tau_{WIN}}{d\sigma} = 0.2150 + 147.11q + 10814q^2 \quad (3.15)
\]

These fits are shown as the blue lines in Figure 3.9. The lower panels show the performance at lower values of specific humidity. In the window region (right), which covers a smaller wavelength range, the parameterisation fits the scatter well over a wide range of values of $q$. In the ‘non-window’ regions, the fit is less clean, with much of the scatter above the blue line at larger values of $q$, and below for smaller $q$. 

92
Figure 3.10: Solid black line: Fraction of blackbody emission in the window region as a function of temperature, T. Dashed red line: Fit to this using Equation 3.16.

An alternative scheme can be parameterised by relaxing the approximation of a fixed proportion of blackbody emission in the window region. Figure 3.10 shows how the fraction of blackbody in the window region varies as a function of temperature. This can be approximated by a quadratic relation:

\[ WIN = -0.0967 \left( \frac{T}{100} \right)^2 + 0.6516 \left( \frac{T}{100} \right) - 0.7089 \quad (3.16) \]

Alternative parameterisations can then be obtained for optical depths calculated from the SBDART output:

\[ \frac{d\tau_{LW}}{d\sigma} = 1.35 + 191.925q^1 \]
\[ \frac{d\tau_{WIN}}{d\sigma} = 0.2914 + 146.1767q + 10433q^2 \quad (3.17) \quad (3.18) \]

Figure 3.11 shows the fits achieved by this parameterisation. The performance in the window region is similar to the fixed window result. In the non-window regions
the dependence of optical depth on humidity now follows a clearer pattern, but the scatter remains wide. The fit still lies below the scatter at high \( q \), although at low \( q \) the improvement can be seen, with the parameterisation following the data more closely.

The first scheme has the advantage of simplicity and needing fewer calculations to be performed. The second allows the effect of the approximation of fixed window fraction to be assessed and provides a better fit, particularly at lower humidity. In the next section the heating rates calculated using both schemes will be compared, and the climates produced when MITgcm is run with each will be presented.

3.4 Incorporation in the MITgcm

For comparison with Figure 3.2, Figure 3.12 shows the heating rates calculated for the temperature and humidity profiles shown in Figure 3.1 using the longwave (Equa-
Figure 3.12: (Left) longwave and (right) shortwave heating rates, K/day, for the reference profiles shown in Figure 3.1 based on the parameterisations presented in this chapter and used in the MITgcm (Equations 3.14, 3.15 and 3.4).

gations 3.14 and 3.15) and shortwave (Equation 3.4) parameterisations. The shortwave scheme, shown in the right hand panel, predicts the heating rates well throughout the atmosphere, although the peak in warming over the equator is overestimated by ~0.2 K/day. To allow comparison with the results acheived by the new scheme, Figure 3.13 shows the shortwave heating rates calculated using the Lacis and Hansen [1974] parameterisation. The structure of the SBDART heating rates is again captured well, with heating decreasing with increasing latitude, and peak heating rates in the correct location, although of slightly weaker magnitude.

Figure 3.14 shows the fractional differences from SBDART associated with the heating calculated using the new scheme (left) and using that of Lacis and Hansen [1974] (right). The largest differences between the new parameterisation and SBDART are at the top level and towards the poles, where the magnitude of heating is weakest. The heating rates in the tropics are in general overpredicted. However, comparing with the right hand plot suggests that the new fit does perform better than that of Lacis and Hansen [1974]. The older scheme consistently underpredicts heating throughout
Figure 3.13: Shortwave heating rates, K/day, for the reference profiles shown in Figure 3.1 based on the parameterisation of Lacis and Hansen [1974].

Figure 3.14: Fractional difference from SBDART shortwave heating of (left) the new shortwave parameterisation and (right) that of Lacis and Hansen [1974].
Figure 3.15: Shortwave heating rates, K/day, for the alternative scheme parameterised using a polar humidity profile (Equation 3.5).

the troposphere, with the differences increasing higher in the atmosphere. The new parameterisation has been designed specifically for ease of inclusion in the MITgcm, and performs well compared with this previous simple scheme.

As discussed in Section 3.2, the shortwave fit used in Figure 3.12 was based on parameterisation to a tropical atmospheric profile. Figure 3.15 shows the heating rates evaluated when a polar profile is used (Equation 3.5). This fit strongly overestimates the tropical heating rates, while higher latitudes now also receive a slight increase in warming. Overall the fit is not an improvement, and the best choice still seems to be to focus the parameterisation on reproducing the low latitude heating rates.

In the longwave, the key features of the SBDART heating rates (Figure 3.2) are reproduced: stronger cooling at near surface levels, a cooling peak over the equator at around 300hPa, cooling rates of between -1 and -1.5 K/day throughout most of the extra-tropics and between -1.5 and -2 K/day in the tropics. Returning to Figure 3.12, it can be seen that at upper levels the scheme performs less well, failing to capture the regions of stronger cooling at higher latitudes. Additionally, the magnitude of the areas of strong cooling are overestimated. However, compared with the results of the
Figure 3.16: Longwave heating rates, K/day, for the alternative scheme allowing for varying blackbody emission by the window region (Equation 3.16).

existing schemes (Figure 3.3) the heating rates are much improved.

In Figure 3.16, the results of the second scheme presented in Section 3.3, where the fraction of blackbody emission in the window is allowed to vary with temperature, are shown. This fit performs marginally better at upper levels, predicting stronger cooling at 50° latitude, and reduced cooling over the equator. The peak cooling values are closer to those of SBDART. However, the peak in cooling over the equator has been shifted lower in the atmosphere than that of the SBDART output. The upper level region of reduced cooling in SBDART has here become a region of warming, with a stronger meridional gradient than that of Figure 3.12. The zonal wind is in thermal wind balance with the temperature distribution, and is known to be sensitive to heating perturbations, particularly around the tropopause [Lorenz and DeWeaver, 2007]. This overprediction of gradient may therefore result in feedbacks onto the jets.

Figure 3.17 shows the zonal mean temperatures and zonal wind resulting from inclusion of the shortwave and fixed window fraction longwave schemes (Figure 3.12) in the MITgcm. Comparing with the multi-model mean of the APE [Williamson and Coauthors, 2012], Figure 1.1, the model climate in this configuration is much improved
Figure 3.17: Zonal mean zonal wind speed, m/s, (shading) and temperature, K, (contours) from the MITgcm when run with the radiation scheme presented here using the fixed longwave window fraction.

from that of the current MITgcm radiation schemes shown in Figure 3.4. The superrotation over the tropics is no longer as strong, and the surface winds have strengthened. The temperature structure is now very similar to that of the APE. At upper levels, both show a minimum over the equator at the tropopause, higher stratospheric temperatures at around 50° latitude, and cooler temperatures again over the pole. These temperature gradients are in thermal wind balance with the jets, and so are important in determining the jet strength and position. A key difference between the new MITgcm climate and the APE climate is the jet structure; the APE atlas has only a single jet at around 30° latitude, while in the MITgcm this has begun to split, with a midlatitude, eddy driven jet over the peak in surface wind speed.

The effect on the model climate of allowing a temperature dependent window fraction is shown in Figure 3.18. Figure 3.16 showed that this scheme better predicted some of the features of the SBDART longwave heating rate distribution, but did introduce some stronger heating gradients in the tropics. In the lower atmosphere, the resulting climate is broadly similar to that of Figure 3.17, with stronger surface winds
than those when the Frierson et al. [2006] and Byrne and O’Gorman [2013] schemes were used. Again, a double jet structure is found. The key difference arises from the upper level temperature structure, where the increased warming gradient has resulted in a small peak in temperature over the equator. This is linked to superrotation, as was seen in Figure 3.4. This is not a general feature of the APE models, and may affect wave behaviour at low latitudes. In order to have a working climate comparable with that of previous studies, the parameterisation of longwave radiative transfer assuming a fixed fraction of blackbody emission in the window region has been chosen to be included in the MITgcm for the experiments in the following chapters.

3.5 Conclusions

An idealised radiation scheme, consistent with the schemes currently implemented in the MITgcm, has been produced to allow simple analysis of radiative feedbacks on atmospheric dynamics. The parameterisation presented uses only three bands to cover the entire spectrum, one for the shortwave, one for the infrared window region, and one
for the remainder of the longwave. The transmission coefficients calculated are used to evaluate fluxes using the two stream approximation. The parameterisation provides a good approximation to the magnitude and structure of radiative heating by water vapour, with both the longwave and shortwave schemes a considerable improvement over the heating rates of the current model. The shortwave scheme performs better than the longwave scheme, as this is a smaller spectral range, and, for a clear sky, behaviour in this region is much simpler than in the longwave. However, the longwave scheme does reproduce the approximate spatial structure of the SBDART heating rates. Importantly, the optical depths calculated with the new radiation scheme result in a temperature and velocity structure that better resembles that of previous aquaplanet studies, and the dynamics seen in the real world.

The parameterisation is adequate for clear sky conditions and where water vapour is the key gas to be considered. For cloudy conditions, a simple parameterisation of shortwave scattering could be incorporated, for example following Lacis and Hansen [1974]. The longwave scheme provides a good compromise between simplicity, speed and reliability, and the heating rates produced are a significant improvement on those from other simple approximations of longwave optical depth. Possible improvements that could be made by allowing the fraction of blackbody emission in the window region to vary with temperature have been investigated. However this was found to result in strong equatorial superrotation, and so the simpler scheme has been chosen for inclusion in the model. Evidently parameterisation of further spectral regions would increase accuracy in all bands, but at the cost of computational speed.

This simple scheme could be useful for a range of simulations, for example idealised paleoclimate simulations [e.g. Ferreira et al., 2011], or investigations into land-ocean warming contrasts [e.g. Byrne and O’Gorman, 2013]. Temporal and zonal averages of the transmission coefficients when the model is run to equilibrium can be used in place
of analytically prescribed optical depths, [e.g. Frierson et al., 2006], providing a more realistic distribution of radiative processes.
4

Effects of latent heat release on the zonal mean state

In this chapter the results of experiments in which the saturation vapour pressure of the atmosphere is varied for fixed optical depths (see Section 2.2) will be presented. The changes to the diabatic heating and consequent changes to static stability, meridional overturning, the momentum budget, and wave activity will be described and interpreted. These effects will then be compared with changes resulting from increasing the SST in the tropics in the *tropheat* experiment. Finally, the changes in eddy length scale in midlatitudes will be presented for each experiment and compared with the Rossby deformation radius and Rhines scale.
4.1 “Moisture content” experiments

As the moisture content of the aquaplanet is increased, the diabatic heating of the atmosphere will change, with increased contributions from large-scale latent heat release and convective heating. Latent heat release is known to result in increased atmospheric stability (see Section 1.3.1). The altered temperature structure, and dynamical feedbacks resulting from this, lead to changes to the mass streamfunction, the zonal momentum budget, and the propagation of eddies. This section describes these changes and attempts to interpret the effects of increased moisture content on the circulation.

4.1.1 Diabatic heating effects of moisture

The primary effect of altering the moisture content of the atmosphere is to introduce a contribution from latent heat release into the atmospheric heat budget (Equation 1.6). The MITgcm models latent heat release from large-scale condensation and moist convection. Changes to these processes will affect the circulation by both changing the atmospheric stability and providing an additional energy source to the atmosphere.

Figure 4.1 shows the variation of the zonal mean diabatic heating rate, $\frac{\partial \theta}{\partial t}$, as humidity increases. In all experiments there is net warming in the tropics and near the surface, and net cooling at higher latitudes. As the atmosphere moistens, the heat released in the tropics intensifies, while the near surface warming contracts to progressively lower layers. The region of strongest cooling shifts from the midlatitudes towards the subtropics.

Changes to the zonal mean diabatic heating are in equilibrium with changes to both the advective heat transport by the zonal mean circulation, and by eddy heat fluxes. The shading in Figure 4.2 shows the contribution to the heat budget by the eddy terms
Figure 4.1: Zonal mean of total diabatic heating rates $Q$, K/day, for the experiments where humidity is varied. Contour interval is 0.5 K/day; grey line indicates zero contour.
in Equation 1.6: \(-\frac{\partial v'\theta'}{\partial \phi}\) and \(-\frac{\partial \omega'\theta'}{\partial \phi}\). The arrows show the eddy heat flux, \((v'\theta', \omega'\theta')^*\), allowing the contributions of the meridional and vertical fluxes to be seen. Comparing with Figure 4.1, it can be seen that in the midlatitudes, where the eddy heat fluxes are strongest, eddy heat flux convergence has an important role in balancing the diabatic heating. In the tropics, the contribution is weaker, with more of the diabatic heating balanced by zonal mean advection. The general picture is of poleward transfer of heat from lower to upper levels, likely related to poleward ascent of warm air in midlatitude storms, for example on warm conveyor belts (see Chapter 5 for further discussion of storm systems in the model.) As humidity increases, the magnitude of the heat convergence increases, showing that eddies are responsible for increasing amounts of heat transport. The 150 experiment shows an additional heating dipole in the tropics, found to result from the \(-\frac{\partial \omega'\theta'}{\partial \phi}\) contribution. This experiment behaves differently to the others in several respects discussed throughout the chapter, including relatively deep equatorial easterlies, which may relate to the low latitude vertical wave activity seen in Figure 4.2. Eddy behaviour in the experiments will be discussed in more depth below.

Looking at the individual sources contributing to the diabatic heating allows changes to this to be interpreted (see Section 2.1.2 for a summary of the parameterisations of these processes in the model). In the dry experiment the only diabatic heating terms are radiative and diffusive heating (Figures 4.3 and 4.4). At the surface, radiative cooling due to the air-sea temperature contrast compensates for diffusion of sensible heat in the boundary layer. At higher levels, the cool temperatures mean that short-wave heating dominates the radiation budget, resulting in net radiative warming in the tropics.

*Note that \(v'\theta'\) has been scaled by a factor of \(180/\pi\), giving the flux in \(^\circ/\text{s}\), so that the arrows are consistent with the convergence when plotted.
Figure 4.2: Contours show the contribution to the heat budget of the sum of the eddy heat flux convergences $-\frac{\partial \omega'}{\partial p}$ and $-\frac{\partial \omega'}{\partial p}$. Contour interval is 0.5 K/day. Arrows indicate the eddy heat flux: $(\nu' \theta', \omega' \theta')$. 
Figure 4.3: As Figure 4.1 but for radiative contributions to heating rates, K/day. Contour interval is 0.5 K/day.
Figure 4.4: As Figure 4.1 but for diffusive contributions to heating rates, K/day. Contour interval is 0.5 K/day.
As humidity is increased, contributions to the diabatic heating from convective heating (Figure 4.5) and large-scale latent heat release (Figure 4.6) strengthen. In general large-scale condensation is weaker than convective heating and is more prevalent in the midlatitudes, where the atmosphere becomes saturated in ascending air masses in storms. The convective heating is strongest in the tropics, where the atmosphere is most humid. The strength of this tropical heating increases with moisture from the dry to the 125 runs. At this point the tropics become saturated at a grid-box scale, and the latent heat release peaks then shift over to the large-scale condensation scheme in the 150 run, so that the combined effect of the two schemes is consistent between experiments (Figure 4.1).

Latent heat release increases the static stability of the atmosphere (see Section 1.3.1). As a result, the boundary layer becomes shallower (Equation 2.4), so that diffusive heating occurs over progressively fewer pressure levels. In the wetter experiments a dipole emerges in the diffusive heating in the midlatitudes, indicating a downward diffusive flux of heat. Additionally, as moisture content increases, the warming of the atmosphere results in stronger radiative cooling, particularly at lower latitudes and higher levels (Figure 4.3). These changes to the heat budget as humidity increases will directly alter the temperature structure of the atmosphere, which in turn will affect the circulation.

4.1.2 Temperature and Stability

Figure 4.7 shows the zonal mean potential temperature for each run. In the dry run, the tropics can be seen to be close to dry stability, $\frac{dw}{dp} = 0$. As humidity increases, the tropics are warmed by latent heat release (see Figure 4.1), and the static stability is expected to follow a saturated adiabat (see Section 1.3.1). Figure 4.8 shows PDFs of the difference between the 500 hPa lapse rate and the saturated adiabatic lapse rate.
Figure 4.5: As Figure 4.1 but for convective contributions to heating rates, K/day. Contour interval is 0.5 K/day.
Figure 4.6: As Figure 4.1 but for large-scale latent heat release contributions to heating rates, K/day. Contour interval is 0.5 K/day.
Figure 4.7: Zonal mean potential temperature, K, for the experiments where humidity is varied.
for each experiment in the tropics, i.e.

\[-\frac{d\Gamma}{dz} - \Gamma_s \equiv \Gamma - \Gamma_s \quad (4.1)\]

where \(\Gamma_s\) is computed following Equation 1.28. Lower magnitude \(\Gamma\) corresponds to a more stable profile. In the PDFs, negative values therefore indicate profiles more stable than moist neutrality.

The PDFs confirm that, for each value of \(e_{a0}\), tropical static stability approximately follows a saturated adiabat, with the most probable values generally close to zero, indicating moist neutrality. In the dry and 010 experiments, the PDFs have peaks for \(\Gamma - \Gamma_s\) less than zero. This is due to the presence of radiative heating in the tropics in these experiments, as discussed in Section 4.1.1, which acts to further stabilise the
atmosphere. However, the overall trend is an increase in static stability, approximately following moist neutrality. Because the SSTs are fixed for all experiments, as stability increases, the result is a descent of isentropes in the tropics so that the equator-pole temperature contrast also increases.

Notably, in the 150 experiment, the PDF indicates atmospheric profiles less stable than moist neutrality. Looking at Figure 4.7, it can be seen that close to the surface in this experiment, static stability is particularly low. This appears to be due to the switch to latent heat release by large-scale condensation rather than convection. In the subtropics, the air close to the ground quickly becomes saturated at a grid box scale. The resulting heating strongly warms the lower levels, destabilising the atmosphere. Less moisture is then carried to higher levels. The low stability leads to feedbacks with the diffusion and convection schemes, where low level heating and cooling intensify (see Figures 4.4 and 4.5). The change in physics means that results from this experiment may have to be treated with caution, but it is nonetheless interesting to see the behaviour of the model at this high moisture content.

While in the tropics and at the poles the stability approximately follows saturated and dry adiabats respectively, the midlatitudes tend to be above moist stability. As was outlined in Section 1.3.1, Juckes [2000] proposed a simple scaling relation, relating the bulk moist stability of the midlatitude atmosphere to the temperature gradient across the storm track. Although this theory appears consistent with both reanalysis data [Juckes, 2000] and model data [Frierson et al., 2006; Frierson, 2008], results from Schneider and O’Gorman [2008] suggested that one of the key assumptions of the theory, moist neutrality as a lower bound on stability, does not hold.

Figure 4.9 shows PDFs of the difference between the 500 hPa lapse rate and the saturated adiabatic lapse rate for each experiment over the midlatitude storm track using Equation 4.1. For each experiment, almost the entire distribution has negative
Figure 4.9: PDFs of the difference between the lapse rate and saturated adiabatic lapse rate, K/km, in the storm track at 500 hPa. The storm track is approximated as a 30° latitude band centred over the latitude of peak $\frac{\overline{\omega}}{\eta}$ (see Table 2.2).

$\Gamma - \Gamma_s$, and a cut-off at approximately zero. The spread is generally wider than was seen in Figure 4.8, indicating a range of profiles more stable than moist neutral. These PDFs demonstrate that, in the MITgcm, moist neutrality is a good approximation to the lower bound of static stability. The cut-off is strictest in the drier models, where the lower bound is closer to dry neutrality, below which the profile will be absolutely unstable and strongly damped. As humidity increases the distribution begins to extend to above zero, indicating that a small number of profiles below moist neutral do exist. However, overall it appears that the approximations of Juckes [2000] are applicable, and that moist convection is important in setting the midlatitude stability.

Figure 4.10 shows an equivalent plot to Figure 1.4 [from Juckes, 2000] for the MITgcm runs. In calculating the meridional temperature difference across the storm
track, the difference is taken at the 500hPa pressure level from 20° equatorward to 20° poleward of the storm-track latitude (see Table 2.2). In calculating the bulk stability, the tropopause height at the storm-track latitude is used. It is hoped that adapting the latitudes and the levels looked at in this way will help to account for the differing climates of the experiments.

The resulting plot is qualitatively similar to the Juckes study, with an approximately linear proportionality, aside from the 150 data point which lies above the rest of the scatter. This is consistent with the idea that the bulk moist stability is determined by the temperature variance in the storm track, with moist neutrality as a lower bound. An increased temperature gradient across the storm strack, and so increased baroclinicity, results in more vertical heat transport by eddies, leading to a more statically stable atmosphere. The slope of the proportionality relates to the efficiency of conversion of the meridional temperature gradient to temperature variance. In the 150 experiment case, a low bulk stability is observed compared to the meridional temperature gradient. This is likely related to the destabilisation by large-scale heating observed in this simulation, as noted above.

4.1.3 The meridional overturning circulation

Having explained how moisture affects the thermal structure of the atmosphere via the heat budget, the resulting changes to the circulation will now be discussed. To link this to the above discussion of the changes to heating rates and potential temperature, the meridional overturning circulation of the atmosphere in isentropic coordinates will first be discussed [e.g Held and Schneider, 1999].

The thermodynamic equation (Equation 1.6) can be expanded as:

\[
\frac{D\theta}{Dt} = \frac{\partial \theta}{\partial t} + \vec{\nabla}\cdot \vec{h} + \omega \frac{\partial \theta}{\partial p} = Q
\]

(4.2)
Figure 4.10: Meridional temperature difference across the storm track vs bulk moist stability.

In steady state, this can be interpreted as a cross-isentropic wind being driven by diabatic heating, $Q$, which can be visualised by looking at the meridional mass transport in potential temperature coordinates, shown in Figure 4.11. The latter is evaluated using [Pauluis et al., 2010]:

$$
\Psi(\theta, \phi) = \int_{-\infty}^{\theta} \left[ \int_0^{2\pi} \int_0^{p_0} v\delta(\theta' - \theta'_0)acos\phi \frac{dp}{g} \right] d\theta'
$$  \hspace{1cm} (4.3)

In the drier models, diabatic heating and cooling are relatively weak (Figure 4.1), with some warming close to the surface and in the tropics, and cooling elsewhere. This is reflected by the overturning circulation in Figure 4.11, where for the drier models, following a contour, potential temperature increases slightly in the tropics due to radiative heating, then decreases weakly with latitude due to radiative cooling. The return flow along the surface is warmed by diffusive heating to match the SST. Contours returning below the SST correspond to the flow within the boundary layer, where stability is low, associated with movement of colder air from higher latitudes over warmer ocean.
Figure 4.11: Zonal mean meridional mass transport in potential temperature coordinates, $10^9$ kg/s. Red line shows the SST.
As humidity increases, latent heat release results in stronger cross isentropic flow in the tropics. The increase in radiative cooling discussed above in Section 4.1.1 is reflected by the ‘descending’ branch of the mass transport cell in the subtropics. Latent heat release is sometimes thought of as a heat source for ascent, amplifying the motions of an ascending air parcel. However, comparing Figures 4.7 and 4.11, it can be seen that while cross isentropic flow in the tropics increases with humidity, static stability also increases. This is reflected by the meridional mass transport in pressure coordinates, shown by the contours in Figure 4.12. Here, the changes to the Hadley cell appear small compared with the changes to the overturning seen in Figure 4.11, particularly in the wetter experiments. As air parcels ascend and moisture condenses, latent heat is released to the environment, increasing static stability, rather than further energising the ascent.

From Figure 4.12, it can be seen that as humidity increases the Hadley cell strengthens, broadens, and deepens. The Ferrel cell in contrast does not follow a systematic pattern, strengthening between the dry and the 050 experiments, then weakening again, reflecting reduced midlatitude eddy activity (see Section 4.1.5, below). Comparing the changes to the Hadley cell with the predictions of the theories reviewed in Section 1.3.2 helps to relate the effects of the altered thermal structure of the atmosphere to the changes in circulation. The left hand panel of Figure 4.13 shows the cell height predicted from Equation 1.34, compared with the tropopause height evaluated using the WMO definition (see Section 2.2). Note that the modelled heights are estimated to the nearest pressure level, with a spacing of 40 hPa. The error bars on Figure 4.13 show the level spacings for each run in order to give an idea of the accuracy of the recorded model heights. The figure is therefore intended to show trends rather than accurate predictions. Even accounting for the low accuracy of the heights recorded from the model, the estimate does not fit the data well, with scatter lying as far as
**Figure 4.12:** Shading shows the zonal mean zonal wind speed, m/s, contour interval 5ms$^{-1}$. Black contours show the zonal mean meridional mass transport, 10$^9$kg/s. Note that contour interval for Hadley cell is 30(10$^9$kg/s), while for Ferrel cell is 10(10$^9$kg/s).
Figure 4.13: Left: Hadley cell heights predicted by Equation 1.34 compared with modelled height, with error bars indicating uncertainty on model values due to vertical resolution. Right: Hadley cell extents predicted by the theories in Section 1.3.2 compared with modelled extents. Black crosses indicate the latitude estimated from Equation 1.36. Green crosses are the latitude where the ratio $(p_s - p_e)/(p_e - p_t)$ (see Equation 1.37) exceeds 1. The dotted line indicates a 1:1 match between model result and theory. Symbols identify individual experiments (see Table 2.1 or key on Figure 4.10 above).

3km from the 1:1 line (e.g. the dry and 125 points). However there does appear to be a roughly linear proportionality, suggesting that the lapse rate and emission height are relevant in determining tropopause height. Both terms in Equation 1.34 increase with humidity, with the decrease in lapse rate, $-\frac{dT}{dz}$, as humidity increases balancing the increase in emission temperature as the atmosphere warms. Assuming the formula is indeed relevant, for fixed SSTs, the decrease in lapse rate with increasing moisture content appears to be the key control on the tropopause height, as this also results in a warmer atmosphere and higher emission height.

The right hand panel of Figure 4.13 shows the extent of the Hadley cell, taken as the latitude at which the mass streamfunction (Figure 4.12) changes sign, versus the predictions of the theories described in Section 1.3.2. Of the two estimates, the black crosses, representing the theory from Frierson et al. [2007b] (Equation 1.36) appear to give the better fit to the data. As for the height predictions, while there seems to be a linear proportionality between model result and theory, the scatter does not
lie on the 1:1 line, with the cell width overestimated at lower moisture contents, and underestimated for moister atmospheres. The proportionality does however suggest that the physics involved in this theory is relevant to the cell width and that the Hadley cell might be approximated as extending to the latitude at which the shear of the resulting zonal jet becomes baroclinically unstable.

The estimates of the latitudes where the vertical wave fluxes become deep enough to reach the upper troposphere, calculated using Equation 1.37 [Korty and Schneider, 2008], are indicated by the green crosses. This method gives plausible estimates for the dry, 025, 050 and 075 experiments. For the 010 and 150 runs, the estimates are very low (7° and 19° respectively). For the remaining experiments the estimates are instead too high: in the 100 run the cell is predicted to extend to 79°, while in the 125 run the ratio $(\bar{p}_s - \bar{p}_e)/(\bar{p}_s - \bar{p}_l)$ does not exceed 1. It is unclear why the estimates might be successful for some experiments and not for others, with no clear dependence on moisture content. Overall, this theory does not appear to give a successful prediction of the modelled cell extent. However, Equation 1.37 is a function of near surface temperature gradients, and with another surface condition, rather than a fixed SST with potential for strong sensible heat fluxes, a different result might be obtained.

From Figure 4.13, it seems that some of the physics involved in the changes to the height and width of the Hadley cell with water vapour content might be understood using the simple theories tested in this section, although the predictions are far from perfect. However, the effect on the strength of the circulation has not yet been discussed. The mass streamfunction presented in this section is defined in terms of the zonal mean meridional and vertical wind speeds, and the controls on its strength are therefore the controls on these windspeeds. The meridional wind speed behaviour can be investigated by analysis of the upper level zonal momentum budget, which will be discussed in the following section. In Section 1.1.2, the TEM formulation of the prim-
itive equations (Equations 1.22 and 1.23) was introduced. This formulation removes the strong cancellation between ascent and eddy heat flux, and combines the effects of eddy fluxes of momentum and heat into the E-P flux divergence, separating out the thermally driven component of the circulation. In Section 4.1.5 the changes to the E-P flux and its divergence as latent heat release increases will be presented.

4.1.4 THE JET STREAMS AND MOMENTUM BUDGET

Figure 4.14 shows the vertically integrated momentum budget, Equation 1.40, for levels higher than 660 hPa.† The strength and width of the cell is reflected by the blue line, which shows the meridional advection of planetary vorticity by the mean flow: $f\bar{v}$. This is balanced by the momentum convergence, $C_{total}$. The dominant contribution to $C_{total}$ throughout most of the atmosphere at most latitudes, and particularly in the drier experiments, is from $C_{eddy}$, demonstrating the importance of eddy activity. As moisture content is increased, the accelerations associated with momentum transport by the zonal mean meridional circulation, $C_{zonal}$ increase in magnitude, and become the dominant contribution to $C_{total}$ at low latitudes.

In steady state, and away from the surface, Equation 1.17 becomes:

$$ \bar{v} \left( f \frac{\partial \bar{u}}{\partial y} \right) = \frac{\partial \bar{u} \bar{v}}{\partial y} $$  \hspace{1cm} (4.4)

In the limit where $f \approx \frac{\partial \bar{v}}{\partial y}$, $\bar{v}$ is therefore independent of eddy forcing. In this case, the upper branch of the Hadley cell is angular momentum conserving, and the circulation will respond directly to thermal forcing, as in the Held and Hou [1980] model. Where this limit does not apply, eddy momentum fluxes are important in determining the cell

†Note that issues with the net torque on the atmosphere were identified in Chapter 2. Internally the momentum budget of the atmosphere appears consistent and so a discussion of the terms involved has been included here. Results may, however, need to be treated with caution, and further work would be required to verify implications for the real-world, see Chapter 6.
Figure 4.14: Vertical integral of the terms in the angular momentum budget (Equation 1.40) for each run for levels higher than 660 hPa. Solid line represents $C_{\text{total}}$, dot-dashed line $C_{\text{zonal}}$, dashed line $C_{\text{eddy}}$, and blue line $f\pi$. Thin lines indicate the latitude of the peaks of the upper level and surface zonal wind speeds.
strength [Schneider et al., 2010]. In the momentum budget, the angular momentum conserving limit is reflected by a balance between $C_{zonal}$ and $f\tau$. At lower moisture content, where $C_{zonal}$ is negligible, the Hadley cell is entirely eddy driven. As moisture increases, and $C_{zonal}$ becomes larger at low latitudes, thermal forcing plays a larger role in setting the strength of the Hadley cell, consistent with the stronger diabatic heating in the tropics seen in Figure 4.1.

The vertically integrated momentum budget is also useful in interpreting the changes to the zonal mean zonal wind, shown by the shading in Figure 4.12. For the dry to 050 runs there is a clear single jet with peak wind speed towards the midlatitudes. As the moisture content of the model increases, the position of this jet shifts poleward only slightly. However, between the 050 and 150 runs, the jet splits into a midlatitude component and a subtropical component.

The thin vertical lines in Figure 4.14 indicate the upper level and surface peaks in the zonal wind speed, giving an indication of the positions of the subtropical and eddy driven jets as the jet splits. In the four drier runs, the upper and lower level peaks in the jet strength lie close to one another and coincide with the peak of $C_{eddy}$. In the four wetter runs, these peaks are well separated, with the lower latitude peak now clearly associated with the poleward branch of the Hadley cell (see Figure 4.12). In these runs the cell is stronger, so that $f\tau$ increases and stronger zonal winds are forced at low latitudes. As the cell widens the eddy driven component of the jet shifts polewards, so that the two components of the jet separate.

4.1.5 Baroclinic Instability and Wave Propagation

From the zonal momentum budget shown above in Figure 4.14, it is clear that the majority of momentum transport, particularly in the extratropics, is associated with eddies, rather than with the zonal mean circulation. As discussed in Section 1.4, ver-
Figure 4.15: Eady growth rate, Equation 1.44, averaged from 900 to 260 hPa.

tical shear of the zonal jet, associated with meridional temperature gradients (thermal 
windshear balance, see Equation 1.12), can result in generation of eddies via baroclinic 
instability. Figure 4.14 illustrates how momentum transport by these eddies acts to 
maintain the zonal jet on which they develop.

The maximum Eady growth rate (Equation 1.44) provides a useful diagnostic of 
regions of baroclinicity in the atmosphere. This is plotted for each experiment in 
Figure 4.15, averaged from 900 to 260 hPa. In the four drier experiments, where only 
an eddy-driven jet was observed, a single peak in the Eady growth rate is seen. As the 
jet separates, two peaks emerge, associated with the shear of the two components of 
the jet ($\frac{\partial \sigma}{\partial z}$ in Equation 1.44). The magnitude of the Eady growth rate decreases with 
mobility content, reflecting the increase in Brunt-Väisälä frequency as static stability 
increases (Figure 4.7). The noise in the growth rate seen at higher latitudes in the 
drier experiments is due to very low static stability in the boundary layer, resulting in 
near zero Brunt-Väisälä frequency.

Comparing this with the E-P flux (Figure 4.16) shows the relation of this baroclinic 
instability to the wave activity in the model. In the boundary layer, low static stability
Figure 4.16: Arrows show the Eliassen-Palm flux (Equations 1.24 and 1.25) for each experiment, scaled following Edmon et al. [1980]. See reference arrows in first panel for scale. Shading shows the E-P flux divergence. Boundary layer data is masked for each run for clarity.
results in very large values of the vertical component of the flux (see Equation 1.25). These have been masked to allow the pattern to be seen between experiments.

The E-P flux is strongest in regions of strong Eady growth rate, and the waves propagate upwards, following the zonal jet. This behaviour is understood from the Charney-Drazin criterion, which states that Rossby waves can only propagate vertically in a background flow that is stronger than their phase speed, and less than some critical speed [Charney and Drazin, 1961]:

\[
0 < U - c < U_c
\]
\[
U_c \equiv \frac{\beta}{k^2 + l^2}
\]  

where \(k\) and \(l\) are the zonal and meridional wavenumbers respectively. The vertically propagating waves must therefore follow the structure of the positive zonal mean zonal wind speed. This explains the lack of vertical wave activity associated with the lower latitude peaks in the Eady growth rate in the four wetter experiments, where the background flow is negative. This criterion also has implications for the vertical component of the E-P flux divergence. As the jet strength increases with height, at some height the critical speed for a given wavenumber, \(U_c\), will be reached, resulting in wave breaking. This is reflected by the decrease in the vertical component of the E-P flux with height, with only the lowest wavenumbers propagating into the stratosphere.†

As water vapour content increases, the region of strongest E-P flux shifts poleward. From the Eady growth rate in Figure 4.15 the region of high baroclinicity can be seen to broaden and shift polewards as the Hadley cell widens. The Hadley cell is in balance with upper level westerlies and lower level easterlies (Figure 4.12). Waves cannot propagate vertically against the low level easterlies, so that the eddy activity

†Wavenumber dependence with height is not shown.
is ‘pushed’ poleward. A reduction in magnitude of the vertical component of the E-P flux with moisture content is also observed. The magnitude of $\overline{v^*}$ increases with moisture content (discussed further in Chapter 5), but the increase in static stability with latent heat release reduces the overall strength of the vertical component of the E-P flux ($\propto \overline{v^*/\theta_p}$).

The contours in Figure 4.16 show the E-P flux divergence. In steady state in the TEM, a non-divergent overturning circulation $\overline{v^*}$ and $\overline{\omega^*}$ is in balance with diabatic heating, friction, and E-P flux divergence, which describes the strength of eddy forcing of the circulation (Equations 1.22 and 1.23). The latter is dominated by the contribution from the divergence of the vertical component of the E-P flux. The flux divergence is strongest in the midlatitudes, where the baroclinic eddies are located. At lower levels, the sign tends to be positive. More eddy activity therefore leaves this region than enters, suggesting that the waves are most strongly generated nearer to the surface. As the waves break in the regions of stronger zonal winds higher in the atmosphere, the divergence becomes negative, with wave momentum transferred to the zonal flow here.

The E-P flux divergence weakens in magnitude with moisture content, indicating a decrease in eddy forcing of the circulation as the strength of the diabatic heating increases. This is due to the reduction in the strength of the vertical component of the E-P flux as stability increases. The decrease in E-P flux divergence as the jet splits is reflected by the decrease in strength of the eddy-driven Ferrel cell (see Figure 4.12).

The horizontal divergence of the E-P flux corresponds to $C_{eddy}$ in Equation 1.40. The balance between this and the advection of planetary vorticity, $f\overline{\pi}$, was seen in Figure 4.14, with $C_{eddy}$ acting to maintain an eddy driven jet in each experiment. In Figure 4.16, the E-P flux can be seen to diverge horizontally in the region of the eddy driven jets, providing additional visualisation of this eddy-mean state interaction.
4.1.6 Summary

The results presented in this section suggest that the key role of water vapour in these experiments is as a strong source of diabatic heating in the tropics. This results in stronger thermal forcing of the Hadley cell, which strengthens and broadens. The equatorward return flow is balanced by low level easterlies, and the region over which eddies may propagate is forced polewards. The jet splits into a component in balance with the upper level branch of the Hadley cell, and a component in balance with the baroclinic wave activity in midlatitudes. Latent heat release in storms, and consequent changes to eddy growth or momentum fluxes in midlatitudes, do not appear to have a strong effect on the zonal mean state.

To investigate the relative roles of zonal mean diabatic heating, and latent heat release in eddies in governing changes in the circulation with moisture content, the next section analyses results of an experiment in which the low latitude SSTs in the 050 experiment are increased, warming the tropics. Restricting the changes in diabatic heating to lower latitudes allows effects on the circulation driven by latent heat release in the tropics to be separated from effects driven from the midlatitudes, for example by changes to eddy growth or propagation.

4.2 Comparison with increased tropical SSTs

This section compares changes to climate produced by increasing tropical SSTs with those from increasing the saturation vapour pressure in the model, discussed above. The 050 experiment is used as a control state for these experiments. For moisture contents above this, the jet begins to split, providing a clear signal to search for in comparing the changes due to increased latent heat release with those due to the SST perturbation. The experiments compared with this control state are the tropheat
Figure 4.17: Comparison of difference in specific humidity, g/kg, between (left) tropheat and 050 experiments and (right) 075 and 050 experiments. Contours show the specific humidity for the 050 experiment for reference.

experiment, in which the model is run with both the SST perturbation shown in Figure 2.3 and with $0.5 \times$ the default value of $e_{a0}$, and the 075 experiment. Changes to the circulation in the tropheat experiment will be driven from the tropics, while in the 075 experiment latent heat release in the midlatitudes may directly affect extratropical eddy momentum fluxes.

Figure 4.17 shows the specific humidity content of both the tropheat and 075 experiments relative to the 050 experiment. In the 075 experiment humidity increases at all latitudes, as would be expected from the increased saturation vapour pressure. In the tropheat experiment, humidity increases in the tropics over the enhanced SST where the atmosphere is warmer, and, aside from regions of decreased humidity associated with changes in the location of ascent and descent in the Hadley cell, remains constant elsewhere.

The effects of the heating perturbation on stability are shown in Figure 4.18. Increased water vapour content between the 050 and 075 experiments stabilises the atmosphere at all latitudes. In contrast, in the tropheat experiment, stability increases
in the tropics where humidity has increased, but the effect is weak compared with the increased saturation vapour pressure. At higher latitudes stability is only slightly increased.

The meridional overturning circulation and zonal jet structure for the *tropheat* experiment are shown in Figure 4.19. A broad Hadley cell and double jet structure are observed, similar to those seen in the four wetter experiments discussed in the previous section (Figure 4.12). The ascending branch of the Hadley cell is narrow compared with the *050* or *075* experiments, reflecting the localised heating in the *tropheat* experiment. This strong, localised thermal driving also results in stronger overturning than is seen in the other experiments.

Comparing the jet structure in the *tropheat* and *075* experiments (Figure 4.20), it can be seen that the effects of the increased forcing from the tropics reproduce the majority of the changes to the zonal wind induced by increased water vapour content. While Figure 4.18 showed that the thermal structures of the two climates are quite different, particularly at higher latitudes, the effects on the zonal winds are very similar,
even outside the tropics. A notable difference is the region of more positive zonal wind speed at low latitudes in the tropheat experiment, which reflects the stronger Hadley cell and increased $f\pi$ term in the momentum budget. However, the extratropical structures of the two experiments are strikingly similar.

The changes in E-P flux are shown in Figure 4.21. The horizontal fluxes undergo similar changes in both experiments, reflecting the changes to the momentum budget with the poleward shift of the jet. However, differences are seen in the vertical fluxes and the flux divergence. These are only weakly affected by the increased tropical SSTs, compared with the effect of increased saturation vapour pressure. This reflects the lack of increase in static stability in the tropheat experiment.

Despite the difference in eddy forcing of the circulation in the tropheat experiment compared with the 075 experiment, the climatologies of the tropheat and 075 experiments are remarkably similar. This indicates that the increased moisture content primarily affects the circulation via stronger zonal mean thermal forcing in the tropics. This drives the changes to the midlatitude eddies, which feed back onto the zonal mean.
Figure 4.20: Comparison of difference in zonal mean zonal wind speed, m/s, between (left) tropheat and 050 experiments and (right) 075 and 050 experiments. Contours show the zonal wind speed for the 050 experiment for reference.

Figure 4.21: Comparison of difference in E-P flux (arrows) and E-P flux divergence (shading), between (left) tropheat and 050 experiments and (right) 075 and 050 experiments. See reference arrows in left hand panel for scale.
Changes to eddy growth or propagation caused directly by latent heat release in storms (e.g. Gutowski Jr. et al. [1992], Booth et al. [2013]), do not appear to significantly affect the zonal mean circulation.

4.3 Eddy length scales

The dominant changes to the atmospheric circulation when water vapour content is increased in the MITgcm have been shown above to be attributable to changes to zonal mean diabatic forcing, with changes to eddy activity arising as a response to this. In Chapter 5, behaviour in individual storm systems as moisture content increases will be investigated in detail. This section looks further at the effect of the changes to the mean state on the eddies, by discussing the changes to eddy length scales with moisture content. In a real world context, the expected length scale of eddies is relevant in understanding the controls on the scale and frequency of storms. In an investigation of the change in eddy length scale in future climates in CMIP3 experiments, Kidston et al. [2010] observed an increase in length scale in all models, suggesting that this a robust feature of the atmospheric response to climate change. Theory indicates that the length scales may be functions of storm track latitude, stability, or eddy magnitude, all of which were shown to be affected by humidity in Section 4.1.

The Eady growth rate discussed in Section 4.1.5 corresponds to the growth rate of the fastest growing mode of the Eady model, and, when compared with the E-P flux, was seen to give a reasonable indication of the regions of high wave activity in the atmosphere. The model predicts that this fastest growing mode has a length scale proportional to the Rossby deformation radius [Frierson et al., 2006]:

\[ L_D = 2\pi \frac{NH_T}{f_0} \]  

(4.7)
According to this theory, the strongest controls on wavelength would come from stability, tropopause height, and latitude (via the Coriolis parameter $f_0$). In their analysis of the CMIP3 experiments, Kidston et al. [2010] identified a correlation between the eddy length scale and the static stability, suggesting that the Rossby radius is the relevant eddy scale for the atmosphere.

Turbulence theory suggests instead that eddy length scale is predicted by the Rhines scale [Rhines, 1976]:

$$L_\beta = 2\pi \sqrt{\frac{v_{EKE}}{\beta}}$$  \hspace{1cm} (4.8)

where $v_{EKE} = \sqrt{\frac{1}{2}(u'^2 + v'^2)}$ is the square root of the eddy kinetic energy. This assumes an inverse cascade of eddy energy to larger scales. Assuming it is not halted by friction or other processes [Sukoriansky et al., 2007], the cascade stops when the vorticity associated with the eddy becomes comparable to that associated with the environmental vorticity gradient, so that the flow in the eddy is affected by the Coriolis force. The length scale is therefore dependent on the Rossby parameter, $\beta$. The key controls on eddy length scale would then be the latitude of the storm track and the kinetic energy of the wave. Barnes and Hartmann [2011] used a barotropic model to investigate the effects of the midlatitude jet location on its characteristics. In a model of this kind, small amplitude Rossby waves are expected to have wavenumbers following $(k^2 + l^2) = \beta/(U - c)$, where $k$ and $l$ are the zonal and meridional wavenumbers, $U$ is the mean flow speed, and $c$ is the phase speed of the waves [e.g. Andrews, 2010]. Barnes and Hartmann [2011] showed that this linear scaling, similar in structure to the Rhines scale, predicted the wavenumbers observed in their nonlinear simulations, with the dependence on $\beta$ resulting in an increase in eddy length scale for a poleward jet shift.
The above length scales can be translated into zonal wavenumbers using:

\[ k = \frac{2\pi \cos \phi}{L} \]  \hspace{1cm} (4.9)

For their experiments with varying moisture content, Frierson et al. [2006] calculated average modelled eddy length scales:

\[ \overline{L} = \frac{2\pi \cos \phi}{\overline{k}} \] \hspace{1cm} (4.10)

\[ \overline{k} = \frac{\int kE(k)dk}{\int E(k)dk} \] \hspace{1cm} (4.11)

In the above, \( E(k) \) is the pressure averaged variance of meridional velocity associated with each wavenumber. These were compared with the Rossby deformation radius and Rhines scale. They found that the mean length scale over the midlatitudes was similar for all experiments. In contrast with the results of Kidston et al. [2010], the Rhines scale provided the best estimate of the observed scales, particularly at the latitude of maximum EKE.

Figure 4.22 shows the wavenumber spectra of \( v' \) as a function of latitude. This was calculated as a 10-year temporal mean of Fourier transforms of \( v' \) snapshots, and averaged over all model pressure levels. The maximum vertically averaged jet speed is marked by the dashed white line. The peak in variance shifts poleward, following the jet, with the associated wavenumber decreasing with moisture. For comparison, Figure 4.23 shows the wavenumber normalised at all latitudes by \( \cos(45)/\cos(\phi) \). At higher latitudes, the same number corresponds to a shorter physical scale. Normalising in this way allows comparison of the length scales at different latitudes, by showing the equivalent wavenumber at 45°.

Overplotted on the spectra in both figures are the wavenumbers predicted by the
Figure 4.22: Wavenumber distribution as a function of latitude for each experiment, calculated as a 10-year average of Fourier transforms of $\psi$ snapshots, and averaged over all pressure levels. The magenta line shows the Rossby deformation radius, while the white line shows Rhines scale. The white dashed line shows the latitude of the peak of the vertically averaged zonal wind speed, indicating the latitude where the jet is strongest.
**Figure 4.23:** As Figure 4.22, but with the wavenumber normalised at all latitudes by $\cos(45)/\cos(\phi)$, indicating the equivalent physical scale at 45°.
Rhines scale (white line) and Rossby deformation radius (magenta line). From Figure 4.22, the Rhines scale appears to give an excellent estimate of the peaks of the spectra at the latitude of maximum variance, consistent with the observations of Frierson et al. [2006]. Additionally, the Rhines scale gives a good prediction of the peak of the spectra at all latitudes above the jet latitude, although it is noted that at the highest latitudes the two scales are similar. A slight increase in $v_{EKE}$ is found as moisture content is increased. However, particularly at higher latitudes, the structure of the Rhines scale wavenumber with latitude is dominated by $\sqrt{\beta}$. At a given latitude, the predicted wavenumber does not change significantly with moisture content. From Figure 4.22 it would therefore seem that the eddy length scale in the extratropics is set by the arrest of an inverse energy cascade by the gradient of planetary vorticity. In particular, the decrease in wavenumber in the storm track as moisture content increases appears to be driven by the poleward shift of the track, and the resulting change in environmental vorticity gradient, rather than by changes to the eddy kinetic energy.

The success of the Rhines scale over the Rossby deformation radius is however more ambiguous in Figure 4.23, particularly at higher latitudes, north of 70°, where both the 45° equivalent wavenumber and Rossby deformation radius wavenumber increase while the Rhines scale predicts a decrease in wavenumber, and equatorward of the peak windspeed where equivalent and Rossby deformation radius wavenumber both decrease. One issue with the use of the Rossby deformation radius in predicting the eddy length scale is the choice of height scale. The estimate used here, derived from the Eady model, uses the tropopause height. If an equivalent estimate is drawn from the Charney model, similar to the Eady model but without a lid at the model top and with the approximation of an f-plane relaxed, the atmospheric scale height, $H = R_{a}T_{0}/g$, is instead used in Equation 4.7. Frierson et al. [2006] tested predictions using both formulae, and found neither to be clearly successful over the Rhines scale, and their
analysis has therefore not been repeated here. However, from Figure 4.23 it appears possible that with some amendment to the scale height used, the Rossby deformation radius could be the relevant scale, and it is therefore difficult to draw firm conclusions on the merits of one over the other.

4.4 Conclusions

The zonal mean climatologies of experiments in which the moisture content of the MITgcm is increased via changes to the model’s approximated Clausius-Clapeyron equation have been presented. Changes to humidity were seen to affect diabatic heating by introducing strong convective heating in the tropics. This alters the temperature structure of the atmosphere, resulting in enhanced upper level radiative cooling as the atmosphere warms, and reduced diffusive heating as boundary layer height decreases. Increased moisture content also affects the midlatitude heat budget, with latent heat release from both grid-scale saturation and convection observed, as well as an increase in eddy heat flux convergence.

Static stability increases with water vapour content. Taking PDFs of the atmospheric lapse rate as compared to a moist neutral lapse rate in the tropics confirmed the tendency for static stability to follow a saturated adiabat at low latitudes. Similar PDFs, produced instead for the midlatitudes, lent support to the theory of Juckes [2000], which suggests that moist convection in the midlatitudes means that moist neutrality can be taken as a lower bound on midlatitude static stability. The effect of waves in enhancing the stability above moist neutral is then expected to be proportional to the temperature gradient across the storm track. Comparison of the bulk moist stability and midlatitude temperature gradient showed results consistent with this theory.

Plotting the meridional mass transport in isentropic coordinates showed the effects
of the increased tropical diabatic heating and subtropical radiative cooling on the circulation. The potential temperature difference between the poleward and equatorward branches of the Hadley cell increases strongly with moisture content. In pressure coordinates, the mass streamfunction shows a deepening, widening, and strengthening of the cell with moisture. The cell height was found to be loosely consistent with radiative-convective equilibrium theories [e.g. Held, 1982, Thuburn and Craig, 1997], while the width was found to be roughly predicted by an amended Held and Hou [1980] theory [e.g. Frierson et al., 2007b], which assumes that the Hadley cell terminates at the latitude at which the shear of the zonal jet predicted by the Held and Hou [1980] model becomes baroclinically unstable. In both cases, the fit is far from perfect, but the proportionality between model result and theory is at least suggestive that these are relevant considerations in understanding the behaviour of the cell height and width.

As moisture content increases the zonal jet divides into subtropical and midlatitude components. Inspection of the momentum budget confirmed that the lower latitude jet is driven by advection of planetary vorticity in the upper branch of the Hadley cell, while the midlatitude jet is eddy-driven. The expansion of the Hadley cell with moisture content causes eddy activity to migrate to higher latitudes, separating the two components of the jet. The momentum budget additionally demonstrated that at low latitudes the advection of planetary and relative vorticity by the mean flow are in balance, so that the circulation here is in the angular momentum conserving limit [c.f. Held and Hou, 1980] and is sensitive to thermal driving. While these results appear internally consistent and logical, issues with the net frictional torque on the atmosphere in the model mean that it would be preferable to investigate the momentum budget further in a different model to strengthen these conclusions.

The Eady growth rate shows a broadening of the regions of high baroclinicity with increasing water content. The E-P flux illustrates the poleward shift in wave activity
as the Hadley cell broadens, with waves unable to propagate vertically in the easterlies found on the lower branch of the cell. The E-P flux divergence decreases with moisture content due to increased static stability. This reflects a decrease in the strength of eddy forcing of the atmospheric circulation, while the strength of thermal forcing increases.

The changes to the model climate as latent heat release increases were then compared with those produced by introducing an increase in SST in the tropics in the 050 set-up. Although the effects on the thermal structure of the atmosphere are different, with the warming and changes to static stability more localised to the tropics with the SST perturbation, the changes to the overturning circulation and jet structure are similar. The Hadley cell deepens and broadens, and the zonal jet splits into subtropical and midlatitude jets, with a very similar structure to the 075 experiment.

A stronger decrease in E-P flux and its divergence was seen between the 050 and 075 experiments than between the 050 and tropheat experiments due to the different static stabilities of the simulations. However, these differences in extratropical wave behaviour do not appear to have any significant affect on the zonal momentum budget, and are not reflected by the meridional overturning or zonal jet structures. These results confirm that the dominant effect of increasing the model water content is to introduce an additional contribution to diabatic heating in the tropics, with most of the differences in circulation driven by low latitude heating, including the changes to the midlatitude momentum budget.

The next chapter will discuss the effects of changes to the heat budget within individual storm systems. However, it is not only local heat release that determines storm structure, and in this chapter the effects of the mean state on eddy length scale were presented. The Rhines scale and Rossby deformation radius were compared with spectra of both the wavenumber and the equivalent wavenumber at 45°, with the latter used to show the actual relative physical scale. From comparison with the
wavenumber spectra, the Rhines scale appeared to give the better prediction of eddy length scale, compared with the Rossby deformation radius. This would imply that the wavenumber is set by a cascade of eddy energy to larger scales, halted by the planetary vorticity gradient $\beta$. The dependence on the $\beta$ parameter means that the Rhines scale is strongly latitudinally dependent, so that wave number decreases as the storm track shifts poleward with increasing moisture content. However, from comparison of both scales with the equivalent $45^\circ$ wavenumber, it appears that with some adjusted scaling the Rossby deformation radius might better capture the modelled behaviour. Results on the relevant scale are therefore inconclusive. Performing similar analysis to that of Kidston et al. [2010], for example looking at the scaling between the length scale and static stability or $\beta$ parameter, might allow stronger conclusions to be drawn.

To summarise, the results of this chapter suggest that the primary control of latent heat release on the mean state is via changes to zonal mean diabatic heating, which results in an increase in tropopause height and an expansion of the Hadley cell. Changes to extratropical eddy behaviour seem to be largely driven by the Hadley cell expansion. The next chapter will discuss how water vapour may additionally have more direct effects on wave structure and eddy energy.
5

Changes to eddy behaviour with increased humidity

In Chapter 4, increased atmospheric moisture content, and the resulting change in latent heat release, were shown to result in significant changes to the zonal mean circulation, and consequent changes to eddy behaviour. This chapter will examine in more detail the effects of latent heat release on midlatitude storms. Firstly, the structure of the storm systems in the different experiments will be presented using instantaneous snapshots of key fields, and pressure-longitude composites centred on cold fronts. Following this, changes to $v'T'$, associated with positive and negative $T'$ anomalies will be presented, with stability in the cold sector demonstrated to play an important role in determining the magnitude of the flux. Finally, the energetics of the
storms will be investigated by comparing the relative amounts of MAPE and EKE in each experiment.

5.1 Storm systems on the aquaplanet

This section gives an overview of the storm behaviour observed in the model simulations before specific changes are discussed in more detail. As described in Section 1.4.1, a useful description of a storm system is that of Browning and Roberts [1994], who discuss the motions of the air masses involved. Figure 1.8 illustrates this conceptual model, showing the movements of the warm and cold conveyor belts and the dry intrusion. Latent heat release in the storm systems can be expected to be concentrated in the warm conveyor belt, a strongly ascending mass of air, moving poleward in the warm sector, ahead of the cold front. Changes to storm strength and development with increased moisture content are generally expected to arise from changes to latent heat transport by this air mass, which is considered to be responsible for most of the transport in the system [Boutle et al., 2011, Booth et al., 2013].

Note: Eddies and Eddy Fluxes

In this chapter, changes to the magnitude and structure of transient eddies and eddy fluxes are illustrated using latitude-longitude maps and PDFs. These will be denoted by primes rather than stars to avoid confusion with the TEM. However, it is important to consider how these may be interpreted. Transient eddy quantities are evaluated by subtracting the temporal mean from the instantaneous field (Equation 1.14). The eddy quantities are therefore dependent on the mean state, which changes when the moisture content of the model is altered (see Chapter 4). A given mean state might be generated both by a symmetric profile of anomalies, for example a sinusoidal profile, or by a more skewed distribution. When drawing conclusions from snapshots
such those in as Figure 5.3, it is therefore important to look at the overall structure of the longitudinal profile, rather than purely at local values.

Additional care must also be taken in interpreting maps of eddy fluxes, for example Figure 5.4. Equation 1.18 gives the zonal mean heat budget in terms of \( \theta \), with eddies contributing to mean advection of heat via cross products of eddy velocity and temperature, e.g. \( \overline{\nu \theta'} \). However, considering the instantaneous local picture, it can be seen that the product of two quantities partitioned following Equation 1.14 will include additional cross terms between eddy and mean state components. Looking only at, for example, \( \nu' T' \) in Figure 5.4, does not show the full picture of how heat is advected locally. This highlights that although maps of quantities such as \( \nu' T' \) indicate how given eddy fields are correlated, and average to produce net fluxes as in Equation 1.6, the instantaneous, local values should not in themselves be assumed to represent a heat flux.

5.1.1 Storm climatology: Snapshots

To give an initial picture of the storm behaviour, Figure 5.1 shows snapshots of the precipitation for each experiment. As might be expected, as moisture content increases, precipitation intensifies. The focus of this chapter is on synoptic scale storm systems in the midlatitudes, but notable changes to the tropical precipitation are also observed. In the drier experiments this occurs in small localised bursts. However, in the wet and 125 simulations a greater proportion of the tropics is seen to be raining at a given moment. In the 150 experiment, the structure returns to intense, local bursts of precipitation. Separating out the contributions to the total precipitation by the convective and large-scale moisture parameterisations, convective precipitation is generally found to dominate in the tropics, with the exception of the 150 experiment, where large-scale condensation is strong (consistent with Figure 4.6). In the drier experiments it can
Figure 5.1: Instantaneous snapshots of the total precipitation rate, kg m$^{-2}$ s$^{-1}$, equivalent to mm/s, for each experiment.
Figure 5.2: Shading shows snapshots of absolute vertical velocity, Pa/s, at 700 hPa. Blue indicates ascent, red descent. Contours show eddy geopotential height, m, at 700 hPa for comparison. Contour interval is 50m. Dashed lines indicated negative values, solid lines positive values.
Figure 5.3: Snapshots of eddy temperature, K, at 820 hPa.
Figure 5.4: Snapshots of $\nu ' T '$, Kms$^{-1}$, at 820 hPa.
be seen that rainfall is organised randomly, with convective precipitation wherever the temperature and humidity profiles become convectively unstable (see Section 2.1.2). As humidity increases, waves of wavenumber 4-5 are seen in the tropics (Figure 4.22), and in the wet and 125 simulations the convective instability and precipitation become organised by the wave. In the 150 experiments, the tropics are saturated at a grid box scale, and the precipitation returns to smaller scale bursts.

In the midlatitudes, changes to the strength and structure of the precipitation are also observed. In the drier experiments weak bands of precipitation are seen to occur around the latitude of the eddy-driven jet. As humidity increases, these bands generally strengthen, with some particularly intense rainfall observed in some individual bands, for example in the snapshot from the wet experiment, a precipitation rate of over \(4 \times 10^{-4}\) mm/s is observed, equivalent to approximately 1.5 mm of rain falling per hour. On days where more than 1mm of rainfall (used to define raining days) falls in the UK, the average intensity is \(~6.7\) mm/day [Kendon et al., 2015], which is of consistent magnitude with the rainfall in the MITgcm simulation. Comparing the precipitation snapshots with those for the vertical wind speed, shown at the 700 hPa level in Figure 5.2, it can be seen that the strongest rainfall is usually associated with strong ascent, resulting in saturation at the grid-box scale. Comparing with the extratropical cyclone climatology of Dacre et al. [2012], shown in Figure 1.7, the model is found to produce ascent of similar magnitude to that found in the ERA-Interim data-set, with values of \(~0.6\) Pa/s (\(~500\) hPa/day) observed. Tropical precipitation is dominated by the contribution from the convection scheme, while the midlatitudes are dominated by large scale saturation.

The contours in Figure 5.2 show the 700 hPa eddy geopotential height. Strong ascent and precipitation are generally found to lie on the eastward flank of a geopotential minimum, where the cold front is located. This structure is again comparable
to that of Figure 1.7. These observations are consistent with these regions of intense precipitation being associated with ascent on the warm conveyor belt. Additionally, a trend towards deeper troughs of eddy geopotential height is observed, with troughs of ~100 m in the drier experiments deepening to troughs of ~200 m in the wetter experiments. That this is not a feature of the individual snapshot shown is confirmed by the upper panel of Figure 5.5, which shows PDFs of the 700 hPa eddy geopotential height for the dry (red) and wet (blue) runs, based on all 10 years of simulation data. This increase in trough depth is consistent with the results of Booth et al. [2013], who observed stronger central pressure minima with increased moisture content in their idealised baroclinic lifecycle experiments.

Figure 5.3 shows $T'$ at 820 hPa. The decrease in wavenumber as the storm track moves poleward observed in Section 4.3 is apparent, with the longitudinal extent of the warm and cold regions broadening with moisture content. The magnitude of $T'$ increases with water vapour content. As was observed in the discussion of the zonal mean state (Section 4.1.2), as the model water vapour content increases over the fixed SSTs, the increased latent heat release in the tropics relative to higher latitudes leads to an increase in the equator-pole temperature gradient. Advection of temperature by the meridional wind on this enhanced temperature gradient results in larger variance of $T'$. Also observed as moisture content is increased is an increase in the strength of the negative values of $T'$ relative to the positive values, indicating a change in symmetry of the temperature field around the mean state. In the dry experiment, significant asymmetry can be seen between positive and negative $T'$ in the wave. By contrast, in the wet experiment, positive and negative $T'$ have both increased in magnitude, with negative values now closer in magnitude to positive. This is further illustrated by the middle panel of Figure 5.5, which shows PDFs of the 820 hPa $T'$ for the dry and wet runs. The distribution of $T'$ in the dry model can be seen to be asymmetric compared
to the wet.

In the midlatitudes, heat is predominantly transported poleward by these extratropical storms. Figure 5.4 shows snapshots of $v'T'$ at 820 hPa for each experiment, indicating the correlation of the temporal eddies about their given mean states. As seen for $T'$, strengthening of $v'T'$ is observed with increased moisture content, in accordance with the increased equator-pole temperature gradient. Another notable feature is the effect of the change in symmetry of the positive and negative $T'$ on the structure of $v'T'$. In the dry experiment, this is seen to be strongest where associated with positive $T'$, or in the warm sector of the storm. However, in the wet experiment, a different picture is seen. Here strong $v'T'$ is associated with both positive and negative $T'$. Across the experiments it can be seen that, as humidity increases, the $v'T''$ in the cold sector increases significantly, with regions of strong $v'T''$ now also associated with colder air. The differences in magnitude of positive and negative $T'$ with moisture content observed in Figure 5.3 suggest that an asymmetry between the warm and cold sector temperatures in the dry model may be responsible for this effect. This is confirmed by PDFs of $T'$ (middle) and $v'$ (bottom) in Figure 5.5. Here it can be seen that the distribution of $T'$ in the dry model is asymmetric compared to the wet, while the $v'$ distributions are relatively symmetric.

As was stated above, changes to the storm dynamics with moisture content are generally expected to be associated with changes to ascent or heat release in the warm conveyor belt. However, looking at the snapshots of $v'T'$, it appears that it is not only the warm sector dynamics that are of interest in understanding the effects of latent heat release on storms. The observed change in the structure of $v'T'$ will be examined in more detail in Section 5.2.
Figure 5.5: PDFs of (top) $\Phi'$ at 700 hPa, (middle) $T'$ at 820 hPa, and (bottom) $v'$ at 820 hPa for the dry (red) and wet (blue) simulations.
Figure 5.6: Composites of diabatic heating, $Q$, K/day (shading) and eddy temperature, K (contours). Composites are averages of variables centred on peaks of the product of relative vorticity and horizontal temperature gradient, $\langle \nabla u T \rangle$, for latitudes within 14° of the storm track latitude as defined in Table 2.2. Grey line marks the zero contour for the diabatic heating.
Figure 5.7: As Figure 5.6, but for the contributions of the different heating terms, K/day, to the total diabatic heating in the wet experiment.
5.1.2 Storm climatology: Composites

While snapshots are useful in providing examples of the storm structures observed in the experiments, it is difficult to clearly see the changes in the average behaviour of the systems as moisture content is increased. Pressure-longitude composites of many storms allow the changes to be more clearly identified, as well as revealing the vertical structure. In producing these composites, the product of the relative vorticity and horizontal temperature gradient, $\zeta \nabla_H T$, at 980 hPa, has been used as an indicator of the positions of the fronts [Sheldon, 2014]. This was evaluated using daily snapshots of $\zeta$ and $T$ for all ten years of each experiment. To account for differences in magnitude of this front diagnostic between experiments, the 95th percentile value was evaluated for each simulation. Peaks with a prominence of at least half of this value were then identified as being significant above the background noise, and used to indicate individual storm systems. Composites were taken for each latitude band within $14^\circ$ of the storm track latitude, as defined in Table 2.2, for both hemispheres. Each composite extends $34^\circ$ in longitude east and west from the peak in the front diagnostic. Where two peaks are within $34^\circ$ of one another, half of the grid points separating the two are assigned to each peak, avoiding double counting of data points. This methodology is similar to that of Messori and Czaja [2015], but with intense values of the front diagnostic used to identify systems to be composited, rather than $\upsilon' H'$. This is because the latter can be associated with both positive and negative $\upsilon'$ and $H'$ (e.g. Figure 1.10), as was seen for $\upsilon'T'$ in the snapshots above.

Figure 5.6 shows composites of the total diabatic heating rate (colours) and eddy temperatures (black contours) for each experiment. Looking first at the composited $T'$ values gives a clearer idea of the thermal structure of the modelled storm systems. Positive $T'$ values are found on the eastward side of the cold front, in the warm sector,
and negative $T'$ is found behind the front, in the cold sector. By definition, the front is associated with a strong horizontal temperature gradient. The general strengthening of the temperature wave suggested by Figure 5.3 can be seen in the composites, with both the warm and cold sectors intensifying with increasing humidity between the dry and wet experiments.

The diabatic heating rate composites reveal significant changes to the heating processes in the storms as humidity increases. As was discussed in Chapter 4, in the dry experiment, the only diabatic contributions to the heat budget are the radiative and diffusive heating terms. In the composites, it can be seen that, at lower levels, diffusive heating dominates. Equation 2.5 shows that this is strongest either in regions of lower stability, where the diffusion coefficient, $K$, is strong, or where there is strong curvature of the potential temperature profile. The diffusive heating acts to damp this curvature, warming colder air and cooling warmer air. This is seen in the dry panel in Figure 5.6, where diffusive heating is associated with cooler $T'$. Radiative processes cool the midlatitudes, with more cooling associated with warmer air.

As humidity is increased, a tongue of diabatic heating is seen to develop along the front, associated with positive $T'$: the warm conveyor belt. This can be clearly seen to intensify with moisture content, as well as extending to higher levels. In Figure 5.7, the contributions to the total diabatic heating for the wet experiment are shown. The heating in the warm conveyor belt is seen to be attributable to large-scale condensation, consistent with the precipitation and ascent snapshots (Figures 5.1 and 5.2). Diffusive heating acts to damp both the warm and cold sectors near the surface, while radiative processes cool the atmosphere at all levels. Convective heating acts to warm the cold sector at lower levels. This is due to the low stability of the cold sector temperature profiles. The contraction of cold sector warming by diffusive and convective processes to progressively lower levels with humidity (Figure 5.6) is attributable to an increase...
in the background stability of the atmosphere.

Further insight into the structure of the storms is given by composites of $\omega$ and $v$. From Figure 5.8 it can be seen that, in all runs, the front is associated with strong ascent flanked by weaker descent. This structure, and the simulated magnitudes of ascent across the front, $\sim 0.25$ Pa/s, are comparable with those observed in similar ascent composites from ERA-Interim reanalysis data [Messori and Czaja, 2015]. It would generally be expected that with increased moisture content, ascent would narrow and strengthen over the front [e.g. Emanuel and Fantini, 1987]. However, in the experiments presented here, the longitudinal width of the ascending region increases with moisture content, reflecting the decrease in wavenumber and slight increase in physical scale as the storm track shifts poleward (see Section 4.3). The strength of ascent decreases slightly with humidity, this is likely related to the increased static stability. Further work is needed to investigate the differences between the modelled changes to the storms with humidity (broadening and weakening of ascent), and the expectations of studies such as Emanuel and Fantini [1987] (narrowing and strengthening of ascent).

In the composites of $v$, Figure 5.9, positive and negative values denote poleward and equatorward moving air respectively. Again, a decrease in wavenumber with humidity can be seen, but aside from this the structure is similar between experiments. Overall the magnitude of the $v$ wave stays relatively constant, with only a slight increase in magnitude with humidity observed. Comparing the composites of $v$ and $T'$ with Figure 1.10, it can be seen that the structures of the modelled storms are comparable to those in the ERA-Interim reanalysis dataset. The tilt of $v$ with height over the front is consistent with the expected structure of a baroclinic system. Despite the lower resolution of the MITgcm simulations, the characteristics of the synoptic scale storms appear to be captured successfully.
Figure 5.8: As Figure 5.6, but for the vertical wind speed, Pa/s.
Figure 5.9: As Figure 5.6, but for the meridional wind speed, m/s.
5.2 RESPONSE OF HEAT TRANSPORT BY EXTRA-TROPICAL WAVES

As discussed in Section 1.4.3, increased moisture content is expected to affect heat transport, potentially by strengthening the warm conveyor belt. This would be reflected by changes to the partitioning of the eddy moist static energy flux (see Equation 1.49) between dry static energy and latent heat with increased moisture content [e.g Frierson et al., 2007a], and by changes to the structures of storm systems. This section will therefore investigate the changes to heat transport in the midlatitudes in more detail, comparing the changes to the moist and dry static energy transport by the wave as humidity increases.

5.2.1 MOIST STATIC ENERGY TRANSPORT

The left hand panel of Figure 5.10 shows the pressure averaged zonal and temporal mean moist static eddy heat flux, $\bar{\nu}T'$, across the latitude of maximum $\bar{\nu}T'$ for each run (see Table 2.2). The black line indicates the total, which is seen to increase with humidity, consistent with the increased equator-pole temperature gradient discussed in the previous section.

Comparing snapshots of $T'$ and $\nu' T''$ (Figures 5.3 and 5.4) in Section 5.1.1, it was noted that, as the model water content increases, the $\nu' T''$ associated with the colder air in the wave increases. To allow changes to heat transport associated with warmer and colder air to be analysed, the red and blue lines show the proportion of $\bar{\nu}H'$ associated with positive and negative values of $T'$, which are here taken as representative of the warm and cold sectors of the extratropical storms. Based on studies on the effect of latent heat release on the warm conveyor belt [Boutle et al., 2011, Schemm et al., 2013], the increase in heat transport would be expected to come via increased latent and sensible heat transport in the warm sector of the storm, and possibly some enhancement
of the storm strength [Boutle et al., 2011, Booth et al., 2013]. As expected, a stronger contribution to the eddy moist static energy flux is given by the warm sector, where the humidity is greater, than by the drier cold sector. However, this is seen to be true for all experiments, including the dry experiment where there is no contribution from latent heat.

The right hand panel of Figure 5.10 shows the contribution of the sensible heat flux to the poleward heat transport. Here it can be seen that the cold sector transport of this quantity actually increases relative to the warm sector transport with moisture content, moving from an approximately 60:40 to 50:50 split, consistent with the changes to $\nu' T'$ observed in Figure 5.4. It seems, surprisingly, that heat transport in the cold sector of the storm has been enhanced by the increased water vapour content. Equivalently, the larger contribution of the warm sector to $\overline{\nu' H'}$ in Figure 5.10 is not only due to an increase in water vapour, but also a weaker dry static energy transport by the cold sector at low saturation vapour pressure.

Current work has focused on exploring the changes in contribution of positive and negative $T'$ air parcels to the eddy heat flux. However, the total eddy heat fluxes also show some interesting behaviour, such as the decrease in sensible eddy heat flux between the wet and 125 experiments. Possible links to the results of previous studies will be discussed in the concluding chapter, along with further work that could be performed to explore this feature.

5.2.2 Sensible heat transport

Different physical processes drive and damp the wave in different atmospheric layers. To aid intercomparison of the experiments, which were demonstrated in Chapter 4 to have very different mean climates, it is useful to divide the atmosphere into two regions. In the first, an approximation to the planetary boundary layer (PBL), interactions with
Figure 5.10: Zonal and temporal mean eddy transport of (left) moist static energy and (right) sensible heat, averaged from surface to tropopause at the latitude of strongest $\bar{v}T'$, as a function of reference $e_{s0}$ fraction. The black line indicates the total, the red and blue show warm and cold sector contributions. Dashed lines in the right hand panel indicate estimated transport, see Section 5.2.3. Shaded areas indicate standard error, assuming data is uncorrelated after 10 days.

Figure 5.11: (left) As Figure 5.10 but for $\bar{T}'^2$ (total not shown). (right) As before but for $\bar{v}'^2$ and with red (blue) line indicating positive (negative) $v'$. Solid lines indicate pressure averages over the free troposphere, dashed lines averages over the boundary layer.
the surface, diffusion and shallow convection are expected to dominate. This layer is defined from the surface to the zonal mean of the PBL height used in the diffusion scheme [Frierson et al., 2006]. The second layer, the free troposphere, extends from the top of the PBL to the tropopause, defined using the WMO definition [WMO, 1957], as in Section 4.1.2. The pressure levels used are shown in Table 2.2.

The left hand panel of Figure 5.11 shows the variance of the eddy temperatures, $\overline{T'^2}$, associated with positive and negative $T'$ in the PBL (dashed) and free troposphere (solid) at the storm track latitude. The PBL can be seen to be responsible for the asymmetry between cold and warm sector $T'$, with a symmetric distribution found in the free troposphere. Equivalent plots of $\overline{u'^2}$ associated with positive and negative $u'$ (right hand panel of Figure 5.11) also show that the positive $u'$ is stronger than the negative in the boundary layer, while in the free troposphere similar magnitudes are found.

### 5.2.3 Mechanism

While the PBL height differs between each experiment, becoming shallower as moisture content is increased, Figure 5.11 shows that $\overline{T'^2}$ and $\overline{u'^2}$ averaged over the PBL model levels vary comparatively little as humidity increases. For all runs, the warm sector is associated with larger $\overline{T'^2}$ than the cold sector, and positive anomaly $\overline{u'^2}$ is greater than negative anomaly $\overline{v'^2}$ in this region. The consistent behaviour between the experiments suggests that a dry mechanism is responsible for the asymmetry.

In the PBL, the temperature and velocity fluctuations in the waves will be damped by diabatic and frictional effects. As previously discussed, the diabatic heating, $Q$, is composed of contributions from radiative heating, diffusive heating, convective heating, and heating due to large-scale condensation. To identify the component responsible for the asymmetric damping, an approximate damping coefficient has been calculated,
averaged over the PBL, by evaluating $\overline{Q \theta'}/(\theta'^2)$ for the warm and cold sectors of each run. The asymmetry appears predominantly driven by the diffusive heating, with the cold sector in the dry run damped by this term by -0.719 day$^{-1}$, compared to -0.556 day$^{-1}$ for the warm sector. In the wet simulation this asymmetry is reduced, with diffusive heating damping the cold and warm sectors by -0.517 day$^{-1}$ and -0.472 day$^{-1}$ respectively. As discussed, the strength of diffusive heating is related to both the curvature of the potential temperature profile and, through the diffusion coefficient, $K$, to the local stability of the atmosphere. Warm air will tend to be cooled and cold air warmed, with the magnitude of heating determined by the diffusion coefficient.

It was shown in Section 4.1.2 that, in each run, the tropics are close to moist neutrality, while the midlatitudes are generally observed to be more stable [Juckes, 2000, Frierson et al., 2006]. In the cold sector of the storm cooler air is moved over warmer water, resulting in a less stable temperature profile. Particularly in the drier experiments, where mean state stability is low, the cold sector can become absolutely unstable, leading to a large model diffusion coefficient and strong warming of the cold air parcel. As humidity increases, the increase in background stability allows larger values of $T'$ to be supported, and the role of the warm and cold sectors becomes more symmetric. Momentum diffusion is also stability dependent, so that similarly motion will be damped more strongly in the cold than warm sector. Evaluating $F\nu'/\nu'^2$, where $F$ is the meridional momentum tendency due to vertical diffusion, confirms this. In the dry (wet) simulation the cold and warm sector $\nu'$ are damped by -1.07 day$^{-1}$ (-1.55 day$^{-1}$) and -0.817 day$^{-1}$ (-1.29 day$^{-1}$).

This stability related mechanism explains the features noted in the $T'$ and $\nu'$ variances in Figure 5.11. However, positive values of $\nu'T'$ will be observed where $\nu'$ and $T'$ are both of the same sign, and so an increase in zonal mean eddy heat flux may be produced by an increase in the magnitude or in the degree of correlation between
these. To quantify how much of the signal over the whole atmosphere is related to damping in the PBL, the contribution of this region to $v’T’$ from the surface to the tropopause has been estimated. Weighting the average of the cold and warm sector sensible heat fluxes in the free troposphere and the individual cold and warm sector fluxes from the PBL by their respective depths provides an estimate of how much of the signal in surface-to-tropopause $v’T’$ is generated by changes to the PBL:

$$\overline{v’T’}_{\text{est cold}} = \int_{p_{\text{pltt}}}^{p_0} \frac{dp}{p_0} \overline{v’T’}_{\text{cold}} + \int_{p_{\text{trop}}}^{p_{\text{pltt}}} \frac{dp}{p_0} (\overline{v’T’}_{\text{cold}} + \overline{v’T’}_{\text{warm}})/2$$

$$\overline{v’T’}_{\text{est warm}} = \int_{p_{\text{pltt}}}^{p_0} \frac{dp}{p_0} \overline{v’T’}_{\text{warm}} + \int_{p_{\text{trop}}}^{p_{\text{pltt}}} \frac{dp}{p_0} (\overline{v’T’}_{\text{cold}} + \overline{v’T’}_{\text{warm}})/2$$

(5.1)

(5.2)

This estimate is shown as the dashed line in Figure 5.10. Taking the ratio of the difference between the estimates and actual values for the dry run, 43% of the total signal is found to be explained by changes to the PBL height and the resulting changes to how diffusive heating damps eddy temperatures. The remainder of the signal would appear to arise from an increase in correlation between same sign $v’$ and $T’$ values.

Close to half of the asymmetry between warm and cold sector values of $v’T’$ is therefore related to diffusion over only a small number of atmospheric layers. The decrease in depth of the PBL with increased moisture content alters the symmetry of the $T’$ wave. These results demonstrate that while changes to warm conveyor belts due to enhanced latent heat release may be the obvious starting point in investigating how increased atmospheric water vapour will affect storm dynamics, more subtle changes to the return flow of colder air, and the associated heat transport, may also be important, and require additional attention.

170
Figure 5.12: Eddy kinetic energy versus mean available potential energy for varying humidity experiments.

Figure 5.13: Ratio of EKE/MAPE as a function of $e_{s0}$ fraction.
5.3 THE EFFECT OF LATENT HEAT ON THE ENERGY CYCLE

Having demonstrated that latent heat release alters heat transport in the midlatitudes, the effects of water vapour on the strength of the systems, in particular their energy, will now be investigated. As discussed in Section 1.4.2, baroclinic waves can be viewed as driven by the conversion of zonal mean available potential energy, associated with unstable temperature gradients, into eddy kinetic energy [Lorenz, 1955]. Evaluating the terms in the atmospheric energy budget in an idealised model similar to that used here, O’Gorman and Schneider [2008b] found a linear scaling of EKE with MAPE (see Figure 1.9). This allows prediction of eddy strength for a given background state, with the relationship seeming to work well in both moist and dry models.

Figure 5.12 shows the values of these diagnostics, evaluated using Equations 1.46 and 1.47, for the MITgcm runs, for comparison with Figure 1.9. In each case these are averaged over the ‘baroclinic zones’ defined as being within 15° of the peak $\overline{v^2 T}$, and integrated to the lowest level of the tropopause. It is clear from Figure 5.12 that this relation is not successful for our data. There are two possible reasons for this lack of correlation: the theory may not apply to these simulations, or there may be some change in conversion of MAPE to EKE as latent heat release increases. The latter effect might not be evident in O’Gorman and Schneider [2008b], where only radiative parameters are changed for a given model.

The equation for MAPE (Equation 1.47) derived by Lorenz [1955], assumes that the distribution of pressure on $\theta$ surfaces is well approximated by the first term of a power series expansion. However, where $\theta$ contours are steeply inclined, this approximation may not be suitable. Inspection of Figure 4.7 shows that particularly at lower levels, and in the dry experiment, where the atmosphere is close to dry neutrality, $\theta$ contours are steep, and so the approximation may not hold. However, even integrating only over
the free troposphere, the MITgcm results still do not give the correlation of O’Gorman and Schneider [2008b].

To investigate the second possibility, that altering the moisture content of the atmosphere has changed the conversion rate of MAPE to EKE, Figure 5.13 shows the ratio of EKE/MAPE for each experiment as a function of reference saturation vapour pressure. This shows a clear increase in the EKE associated with a given amount of MAPE with moisture content. The curve appears to level out at a ratio of approximately 0.7 for the 050 experiment. A significant increase in the conversion rate is also seen between the 125 and 150 data points, which could be related to the high relative humidity in the 150 experiment, reflected by the dominance of the large-scale condensation scheme in this run. It is possible that the results of O’Gorman and Schneider [2008b] could lie in the level region of the conversion graph observed between the 050 and 125 experiments, with extreme values of moisture content altering the conversion rate.

5.4 Conclusions

Snapshots of precipitation and ascent have revealed changes to the storm track structure as moisture content increases. Extratropical precipitation was found to intensify with humidity, with the strongest rainfall associated with strong ascent, indicating a strengthening of the warm conveyor belt. This strengthening is clearly seen in composites of the diabatic heating rate, which show a tongue of warming developing across the front, associated with the core of ascent. This tongue was found to be attributable to large-scale condensation of water vapour as ascending air becomes saturated at the grid-box scale. Also seen in the heating rate composites was a decrease in the depth of warming in the cold sector due to diffusive and convective heating.

Composites of meridional wind across the front showed a slight increase in amplitude
with moisture content, and a decrease in wavenumber consistent with that observed in Chapter 4. Similarly, an increase in the width of ascent and descent was seen in composites of $\omega$, as well as a decrease in strength of the vertical motions. Increased latent heat release has been predicted to result in a narrowing and intensification of ascent [Emanuel and Fantini, 1987, Booth et al., 2013]. It is possible that the increased static stability as moisture increases negates this effect in these large-scale systems. Further investigation would, however, be required to draw firm conclusions.

Looking at snapshots of eddy temperature and $v'T'$, it was seen that in the drier experiments positive eddy temperatures are of greater magnitude than negative eddy temperatures, and are associated with stronger heat transport. As moisture content increases, the magnitude of both positive and negative temperatures increase, reflecting the stronger equator-pole temperature gradient. Additionally, the cold sector strengthens relative to the warm sector, reducing the asymmetry between these and resulting in stronger $v'T'$ values associated with equatorward movement of cooler air.

The behaviour of the cold and warm sector heat transports was investigated further by comparing surface-tropopause averages of $\overline{v' H}$ and $c_p \overline{v'T}$ at the latitude of maximum $\overline{v'T}$ for each experiment. As is generally expected [Boutle et al., 2011, Schemm et al., 2013, Booth et al., 2013], an increase in the latent and sensible heat fluxes associated with the warm sector due to positive feedbacks from latent heat release in the warm conveyor belt was observed. However, it was additionally found that the cold sector heat transport strengthens with moisture content, accounting for 49% of the total change in moist static energy transport, and 61% of the change in sensible heat transport. While in drier models the cold sector provides a proportionally smaller heat flux, this increases with humidity until the cold and warm sectors give a similar contribution.

The strengthening of the cold sector heat transport is partially explained by an in-
crease in the magnitude and spread of negative values of $T'$ with moisture. In the model PBL, cold $T'$ values are strongly damped by diffusive heating, particularly when the temperature profile is unstable. The asymmetric distribution of temperature around the mean results in larger regions of weaker cool $T'$ values versus smaller regions of stronger warm $T'$. As static stability increases with humidity, the modelled PBL depth decreases, allowing more of the atmosphere to support a stronger cold sector. Simply put, at lower levels, bringing warmer air masses over colder water yields a stable profile. By contrast bringing cooler air over warmer air or water, particularly in an environment already close to dry stable, can result in an absolutely unstable temperature profile which is quickly damped, limiting the strength of cold $T'$ that can develop in the storm.

Lastly, to explore the changes in EKE as moisture content is varied, the time mean MAPE and EKE over the storm track were calculated. Previous studies have found a linear relationship between MAPE and EKE, implying that for a given equator-pole temperature gradient and stability, the conversion rate of available potential to eddy kinetic energy is constant [e.g. O’Gorman and Schneider, 2008b]. However, for the experiments presented in this thesis, this was not found to be the case. Plotting the ratio of EKE/MAPE against reference saturation vapour pressure for each experiment showed a smoothly increasing curve that levels out between the 050 and 125 experiments, before increasing again in the 150 experiment. Although further investigation of the mechanism involved is required, this suggests that latent heat release may in some way alter the conversion of MAPE to EKE. The next chapter will discuss ideas for experiments that would allow further analysis of this change.
6

Conclusions

Idealised experiments in which the water vapour content of the atmosphere is varied have been performed with an intermediate complexity GCM, the MITgcm. As the atmosphere warms under climate change, humidity will increase, and understanding the feedbacks that may result from this is important in predicting how the atmospheric circulation can be expected to change. Water vapour affects dynamics by both radiative and latent heating. While radiative water vapour feedback has been extensively researched [Myhre et al., 2013], latent heat feedbacks are complex and difficult to predict. This thesis has therefore largely focused on the latter of these.

The set-up used in the MITgcm for these experiments included simple parametrisations of diabatic heating processes and frictional effects following Frierson et al. [2006] and O’Gorman and Schneider [2008a]. However, the highly idealised parametrisation of radiative heating resulted in mean climates with significant differences from previ-
ous studies [e.g. Neale and Hoskins, 2001]. A new, simple parametrisation of radiative
transfer has therefore been developed and used in the experiments throughout the
thesis.

Experiments where the reference saturation vapour pressure in the model’s approx-
imate Clausius-Clapeyron equation (Equation 2.6) was varied by factors between 0 and
1.5 allowed a wide range of climates with varying water vapour content to be investi-
gated. The effect of latent heat release on the zonal mean state in the model was first
investigated, with significant changes in climatology observed. These were compared
with an experiment in which the SSTs in the tropics were increased, allowing some un-
derstanding of the more important processes in setting the mean state to be achieved.
These experiments were then used to investigate the effects of latent heat release on the
heat transport and energetics of extratropical storms. In Chapter 2, issues with the net
frictional torque on the atmosphere were noted. The results presented may therefore
need to be treated with some caution, although the physical arguments appear logical
and consistent.

In this chapter, the key results of the thesis will be summarised. Following this,
future questions that have emerged from this research, and suggestions for future
work, will be presented.

6.1 Summary of Results

6.1.1 Radiative parametrisation

In Chapter 3, having demonstrated that the simple parametrisations of radiation cur-
rently used in the MITgcm do not compare well with the radiative heating structure
predicted by a fast radiative transfer code, SBDART, a new parametrisation was de-
veloped.
The new radiation scheme uses three bands to approximate radiative transfer over the entire spectrum. Firstly, a shortwave band approximates absorption of radiation with wavelength less than 4µm. Longwave radiation is divided into two ‘types’, an infra-red window region from 8 to 14µm, and all wavelengths outside of this region. For all bands, the two stream approximation is used to evaluate fluxes and heating rates.

In the shortwave, it was found that there was a need to account for the elimination of more strongly absorbed wavelengths at upper levels, resulting in an absorption coefficient that decreases as a function of the optical depth to a given level. In the longwave, the window region was found to be best fitted by a quadratic function of specific humidity, $q$, while the non-window region was fitted as a function of $q^2$.

Comparing the heating rate estimates of the new parametrisation with the predictions of SBDART, the shortwave scheme was found to perform well, with the heating rates calculated accurate to within 20% over most of the atmosphere, compared with the 30-40% difference achieved by a previous scheme [Lacis and Hansen, 1974]. While the longwave parametrisation does not achieve this level of accuracy, the key features of the SBDART heating rates are captured, and the scheme represents a significant improvement over the current implementation [Frierson et al., 2006, Byrne and O’Gorman, 2013].

With the new parametrisation included, the temperature and velocity structures of the MITgcm compare much better with previous aquaplanet studies [Neale and Hoskins, 2001]. While the inclusion of additional bands would improve accuracy, particularly in the longwave, the scheme bridges the gap between fixed optical depths and complex radiative transfer calculations. It is hoped that this parametrisation could be useful for a range of future idealised modeling studies.
6.1.2 Effect of Latent Heat Release on the Mean State

Having developed a radiative parametrisation that provides a more realistic mean state, in Chapter 4 experiments were performed where the optical depths were fixed to the zonal and 10 year temporal mean of those calculated by the radiation scheme for the default reference saturation vapour pressure value, \( e_{s0} \). In these experiments, \( e_{s0} \) was varied by factors of 0, 0.1, 0.25, 0.5, 0.75, 1, 1.25, and 1.5, with 0 resulting in a completely dry atmosphere, and 1 giving the default climate. Comparison with an experiment in which a perturbation to the tropical SSTs was introduced (see Figure 2.3) allowed the relative importance of latent heat release in the tropics, versus changes to eddy propagation in the midlatitudes, in producing the response to increased moisture content to be investigated.

As humidity was increased, significant changes to the model’s mean climate were observed, particularly:

- An increase in equator-pole temperature gradient
- An increase in static stability
- Increases in the width, height, and strength of the Hadley cell
- A poleward shift in the zonal jet, with a split of the jet into subtropical and eddy-driven components seen between the \( \theta50 \) and \( \theta75 \) experiments
- A poleward shift in wave activity, and a decrease in vertical component of the E-P flux
- A decrease in wavenumber and slight decrease in equivalent 45° wavenumber (decrease in wavelength) in the storm track
The changes to the meridional temperature gradient result from stronger latent heat release at lower, compared with higher, latitudes. In the tropics, static stability follows a moist neutral profile, the stability of which increases with moisture content. In the midlatitudes, PDFs comparing the observed atmospheric lapse rates with moist neutrality gave support to the theory that moist convection sets the lower bound on the midlatitude lapse rate, with waves enhancing the stability [Juckes, 2000]. Comparing the midlatitude temperature gradient and bulk moist stability also supported this theory, with a rough proportionality observed.

Theories for the height and width of the Hadley cell were used to interpret the observed changes to these. While providing a poor fit to model results, radiative-convective equilibrium theory [e.g. Held, 1982, Thuburn and Craig, 1997] correlates with the increase in cell height observed with increased water vapour. If this theory can be taken as applicable, over the fixed SST, the decrease in lapse rate with increasing humidity would seem to be a key control on the changes to tropopause height. The cell width was found to be roughly predicted by the amended Held and Hou [1980] theory [e.g. Frierson et al., 2007b], where the cell is assumed to terminate where the shear of the zonal jet of the Held and Hou [1980] theory becomes baroclinically unstable. Again, the fit is far from perfect, but the correlation does suggest that this theory may be applicable to some extent. The theory does not account for latent heat release, suggesting that moist processes may not explicitly complicate the prediction of the cell width, although they do have an effect via changes to the background thermal structure of the atmosphere. Comparing the changes to the Hadley cell with those seen in the *tropheat* experiment supports this, with similar changes to structure observed, despite the difference in mechanism for heating the tropics.

The expansion of the Hadley cell with increased moisture content also results in a shift of eddy activity to higher latitudes. This shift in wave activity as the cell
broadens appears to be responsible for the poleward shift and split of the jet. In the drier experiments, the jet is predominantly eddy driven. The strengthening and increased extent of the poleward branch of the Hadley cell results in the formation of a subtropical jet, in balance with the advection of planetary vorticity at upper levels. These changes to the jet are also seen in the tropheat experiment. In this experiment, midlatitude stability is not strongly altered compared with the 050 experiment, and the E-P flux is not significantly changed. Differences in climate in this experiment are instead driven entirely from the tropics, where the SST is perturbed. The similarity in jet behaviour between this experiment and those with increased moisture content indicates that the altered meridional temperature gradient due to the warming of the tropics is responsible for the observed shift in the jets and eddy activity. Changes to latent heat release in storms (e.g. Gutowski Jr. et al. [1992], Booth et al. [2013]) do not have a significant effect on midlatitude eddy momentum fluxes.

From wavenumber spectra produced from Fourier transforms of v′, the Rhines scale appeared to best predict the observed changes in wavenumber in the experiments. This suggests that the wave activity in the model can be viewed as a cascade of energy to larger scales, with the planetary vorticity gradient, β, stopping the cascade. The β parameter is a function of cosφ, so that as the jet shifts poleward with increased moisture content, the wavenumber decreases. Looking instead at the equivalent wavenumber at 45° reduces confidence in this conclusion, with the shapes of some features of the changes in physical scale, such as the low latitude structure and an increase in wavenumber at high latitudes, better predicted by the Rossby deformation radius, although some scaling factor may be missing in estimating the correct magnitude. Looking at the correlation of modelled eddy length scale with stability and β parameter might allow firmer conclusions to be drawn.
6.1.3 Effect of latent heat release on midlatitude eddies

Looking at the mean state of the experiments has shown that, for fixed SSTs, the predominant effect of latent heat release is to preferentially warm the tropics, where the warmer air holds more moisture. Latent heat release is also associated with midlatitude storms, and a growing area of research has been understanding how feedbacks from moisture affect storm growth and development [Schemm et al., 2013, Booth et al., 2013]. While the mean states of the experiments in this thesis are rather different, they may allow a more complete picture of latent heat effects on the midlatitudes than the eddy lifecycle experiments performed in previous studies, where fixed initial conditions are used. Chapter 5 investigated the changes to extratropical storms in the experiments.

Comparing snapshots of precipitation and vertical velocity, and looking at composites of diabatic heating across fronts, it was seen that ascent on the warm conveyor belt in storms is associated with strong latent heat release and precipitation. The magnitude of the heat released increases with humidity as the storms moisten. Additionally, the geopotential height anomalies increase in magnitude, consistent with predictions of eddy lifecycle experiments [e.g. Booth et al., 2013].

It has been predicted that latent heat release should result in an intensification and narrowing of the ascent in storms [Emanuel and Fantini, 1987]. Here a broadening and decrease in strength of ascent has been observed. This has been attributed to the poleward shift in the storm track, which results in a decrease in wavenumber. The low resolution used in these experiments may mean that smaller scale motion cannot be resolved, so that narrowing is not seen. However, comparing composites of meridional velocity and eddy temperature with those produced from ERA-Interim data [Messori and Czaja, 2015], it appears that the structure of the baroclinic systems is comparable
with the reanalysis, suggesting that the resolution is adequate. The increased static stability as moisture content increases may negate this effect, although this hypothesis requires further analysis.

A key result of Chapter 5 was the observation of a change in structure of the midlatitude waves with moisture content; specifically that as humidity increased, the strength of the cold sector $T'$ increased in magnitude relative to that of the warm sector indicating a change in symmetry of the temperature field variations. While in the dry experiment, roughly 60% of $\overline{v'T'}$ is associated with positive $v'$ and $T'$, as moisture content increased, the cold sector $\overline{v'T'}$ increased to account for 50% of the total eddy heat flux.

A mechanism for this change was proposed and tested. Negative $T'$ is associated with moving colder air over warmer air or water, destabilising the atmosphere. In the drier models, where static stability is low, and dry neutrality acts as a lower bound, destabilisation may result in absolutely unstable temperature profiles, which are strongly damped. As moisture increases, the static stability of the atmosphere increases, with moist neutrality as a lower bound, and movement of cool air is less likely to result in strong destabilisation. In the model, this is reflected by asymmetry between the variance of temperature (meridional velocity) associated with positive and negative $T'$ ($v'$) in the PBL. Diffusive and convective heating damp cold $T'$, with the magnitude of the damping dependent on static stability. Similarly, momentum diffusion damps $v'$, again dependent on stability. While the intuitive picture of how latent heat release will affect storm development is to think of changes to the WCB, these results show that a more thorough understanding of the interplay of moisture, stability and heat release is needed.

Calculating values of EKE and MAPE for each experiment, an apparent dependence of the ratio of these on humidity was found. This is in contradiction with previous
studies, where a linear relationship was observed [O’Gorman and Schneider, 2008b]. However, while previous work varied insolation and longwave optical depth in both dry and moist models, the effect of latent heat release on the energy cycle has not been directly explored. It is possible that water vapour may affect the conversion of MAPE to EKE, as is suggested by the results from these experiments. Ideas for experiments to investigate this further are discussed in the following section.

6.2 Future work

From the above summary, it can be seen that some open questions remain. Ideas for future work to build on the results of the thesis are outlined below.

6.2.1 Further experiments with a slab ocean

One clear outstanding question from this work, which has used an idealised model, is the relevance of the results of these experiments, in which moisture content is artificially varied, to the real world. Repeating the analyses performed here on reanalysis data, or data from more comprehensive climate models, would allow the results presented here to be explored further. A first step could be to repeat the experiments in a model with a Q-flux ocean. For example, the physical arguments presented here for the change in heat transport behaviour of cooler and warmer air parcels are not expected to be dependent on the surface condition, and further work with a more realistic surface condition, and ideally in a model lacking the issues with the angular momentum budget observed in Chapter 2, could allow the conclusions made here to be more firmly related to the real world.

Investigation of some behaviour not discussed in detail here, such as the decrease in sensible eddy heat flux between the wet and 125 experiments seen in Figure 5.10, could also be performed, using a Q-flux model, with this goal in mind. Held and Soden

185
[2006] investigated the changes to the hydrological cycle under global warming using the archive of climate experiments from the Program for Climate Model Diagnosis and Intercomparison (PCMDI) for the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change. Amongst the robust changes observed in these models is a decrease in horizontal sensible heat transport in the extratropics, compensating for an increase in the eddy moisture flux. It is appealing to attempt to relate the decrease in $c_p \bar{v'} T'$ is seen between the wet and 125 experiments to the AR4 result. However, from Figure 5.11 it would appear that it is the variance of $v'$ that is responsible for this behaviour, and that this is therefore unlikely to indicate any compensating behaviour between the moist and dry static energy fluxes in this model set-up. With the fixed SST surface condition, the total poleward heat flux is not constrained to balance the gradient in incident solar and outgoing longwave radiation, so the arguments of Held and Soden [2006] do not necessarily apply here. Investigation with a Q-flux ocean would allow research into this potentially compensating behaviour.

6.2.2 EXTENSION TO THE RADIATION SCHEME

The radiation scheme presented in this thesis has been demonstrated to perform well compared to existing simple parametrisations. However, additional work could allow for a wider range of applications. In particular, the scheme does not account for scattering of radiation in the shortwave. As the Frierson et al. [2006] physics package does not include clouds, with precipitation falling straight out following condensation, these were not considered when developing the radiation scheme. However, cloud radiative feedbacks are a large source of uncertainty in predictions of climate change [Boucher et al., 2013]. An intermediate complexity model including clouds could be of use in interpreting results from more complex model studies of how clouds affect dynamics.

If the Frierson et al. [2006] and O’Gorman and Schneider [2008a] physics scheme
were extended to include clouds, an existing simple parametrisation of scattering [e.g., Lacis and Hansen, 1974] could be added to the radiation scheme that has been developed here. Parametrisation of the longwave effects would need further consideration and discussion, but these could potentially be incorporated by approximating layers of cloud as black, or grey, bodies.

It has also been noted that while the radiation scheme performs well in the shortwave, the longwave component is less accurate. Improvements could be made here by increasing the number of spectral regions described, although this increases the complexity of the parametrisation, and might compromise speed.

6.2.3 The effect of latent heat on the energy cycle

In Chapter 5, the relationship between MAPE and EKE as moisture content is varied was briefly investigated. From previous studies [O’Gorman and Schneider, 2008b], a linear proportionality was expected, so that EKE would be determined by the stability and meridional temperature gradient. However, this linear relationship was not observed, with the ratio of EKE/MAPE smoothly varying with moisture content. It is possible that altering the saturation vapour pressure in the Clausius-Clapeyron equation somehow alters the conversion of MAPE to EKE.

O’Gorman and Schneider [2008b] varied the radiative parameters of an idealised model, in particular insolation and longwave absorber density, to provide a range of meridional temperature gradients and climates. A similar methodology could be applied here. Varying the insolation gradient is expected to only weakly affect the climate when fixed SSTs are used. However, by including a slab ocean, or by varying the SSTs directly, a range of climates could be generated for each value of $e_{s0}$. Comparing the resulting EKE and MAPE for each moisture content would provide further information on whether the ratio of EKE/MAPE can be taken as constant for a given model.
Analysis of the processes involved in the generation and conversion of energy in the atmosphere [Lorenz, 1955] would then allow a mechanism to be identified.

Ambaum and Novak [2014] discuss an alternative perspective for the conversion of baroclinicity into wave activity, viewing the system instead as a nonlinear oscillator. Rather than discussing the growth and decay of eddies, they describe these as forced oscillations, where baroclinicity continually builds up and is released in bursts of heat transport. The mean state is then the centre point of this oscillating system, not the most likely climate of the system, but rather the state about which it oscillates. The time scale of the oscillations would be set by the diabatic forcing, and could change with increased moisture as the magnitude of this forcing changes. This perspective might provide an interesting way to analyse the results of the experiments described above.

In summary, although further research into the role of latent heat release in climate is still required, the results of this thesis show that this strongly influences the circulation. In the zonal mean, the distribution of water vapour is important in determining the distribution of diabatic heating, which in turn controls the thermal structure of the atmosphere. The effects of latent heat release on extratropical storms have been shown here to be complex, and not always intuitive, with changes to stability with increased moisture content potentially important in determining the structure of eddies. Additionally, preliminary results suggest that increased latent heat release may alter the energy cycle, resulting in more energetic eddies for a given mean state. It is hoped that future work, both directly expanding on these results, and potentially comparing with data from more complex GCMs or reanalysis, will allow further development of our understanding of the role of water vapour in the climate system.
References


193


<table>
<thead>
<tr>
<th>Page No.</th>
<th>Type of work</th>
<th>Name of work</th>
<th>Source of work</th>
</tr>
</thead>
<tbody>
<tr>
<td>26</td>
<td>figure</td>
<td>Figure 4.15: Zonal-time average APE multi-model mean and standard deviation zonal wind (u), temperature (t), meridional wind (v), vertical wind (om), specific humidity (q) and relative humidity (rh).</td>
<td>Williamson, D. L., and Coauthors, 2012: The APE atlas. NCAR Technical Note NCAR/TN-484+STR, doi: 10.5065/D6FF3QBR</td>
</tr>
<tr>
<td>33</td>
<td>figure</td>
<td>FIG. 1. Graphical demonstration of the calculations of (a) the maximum meridional heat transport (MHTMAX), (b) ASR*, and (c) OLR* from the CERES annual average data. The x axis is the sine of latitude in all panels. In (a) the zonal average ASR (red line) and OLR (green line) are shown. The blue- (red-) shaded area is the spatially integrated net radiative deficit (surplus) in the extratropics (tropics) and equals the meridional heat import (export) from each region (MHTMAX). In (b) the zonal average ASR is coplotted with the global average ASR; the shaded area equals ASR*. (c) Is as in (b), but for OLR and OLR*. The black dots denote the latitude where ASR9 and OLR9 5 0 in each hemisphere.</td>
<td>Aaron Donohoe and David S. Battisti. What determines meridional heat transport in climate models? J. Climate, 25:3832-3850, 2012. ©American Meteorological Society</td>
</tr>
<tr>
<td>34</td>
<td>figure</td>
<td>Fig. 3.14 Infra-red absorption spectra for six strongly absorbing gases and for the six gases combined, for a vertical beam passing through the atmosphere, in the absence of clouds. Drawn from data supplied by Dr A. Dudhia.</td>
<td>David G. Andrews. An introduction to Atmospheric Physics - Second Edition. Cambridge University Press, 2010.</td>
</tr>
<tr>
<td>39</td>
<td>figure</td>
<td>FIG. 6. The vertical change in equivalent potential temperature against the meridional change in midtropospheric temperature. Symbols as in Fig. 2, (a) with relative humidity $h = 1$ and (b) $h = 0.75$.</td>
<td>M. N. Juckes. The static stability of the midlatitude troposphere: The relevance of moisture. J. Atmos. Sci., 57:3050–3057, 2000.</td>
</tr>
<tr>
<td>41</td>
<td>figure</td>
<td>FIG. 1. Vertical cross sections of annual-mean (a) zonal wind $[u]_an^<em>$ and (b) MMC $[\psi]_an^</em>$. Solid contours are positive, dashed contours are negative, and the zero line (apparent only for $[u]_an^*$) is thicker. For the zonal wind, the contour interval is 5 for positive values and 2 ms$^{-1}$ for negative values ($\ldots$, 4, 2, 0, 5, 10, $\ldots$). For the MMC the contour interval is $2 \times 10^{10}$ kgs$^{-1}$ ($\ldots$, 3, 1, 1, $\ldots$).</td>
<td>Ioana M. Dima, John M. Wallace, and Ian Kraucunas. Tropical zonal momentum balance in NCEP reanalyses. J. Atmos. Sci., 62:2499–2513, 2005.</td>
</tr>
<tr>
<td>44</td>
<td>figure</td>
<td>FIG. 6. Vertical cross sections of the leading terms in the annual-mean momentum budget: (a) MMC associated fluxes $[v][f - (1/\cos \phi)(\partial[u] \cos \phi/\partial y)]_an^<em>$, (b) eddy fluxes $[(1/\cos^2 \phi)(\partial[u^<em>v^</em>] \cos^2 \phi/\partial y)]_an^</em>$, and (c) sum of the two contributions in (a) and (b). Solid contours are positive and dashed contours are negative. The contour interval is $1 \times 10^3$ ms$^{-2}$ ($\ldots$, 1.5, 0.5, 0.5, $\ldots$).</td>
<td>Ioana M. Dima, John M. Wallace, and Ian Kraucunas. Tropical zonal momentum balance in NCEP reanalyses. J. Atmos. Sci., 62:2499–2513, 2005.</td>
</tr>
<tr>
<td>50</td>
<td>figure</td>
<td>Fig. 4. Horizontal composites at (a) 48 and (b) 24 h before the time of maximum intensity, and (c) at the time of maximum intensity. Bottom row: 925-hPa geopotential height (solid lines at 400, 600, and 800 m); system-relative wind vectors, and frontal positions. Middle row: 700-hPa geopotential height (solid lines at 2,800 and 3,000 m); equivalent potential temperature ($\theta_e$, dashed lines at 292, 300, and 308 K) and vertical velocity (omega, filled). Top row: 300-hPa geopotential height (solid lines at 8,600, 8,800, 9,000, and 9,200 m); $\theta_e$ (dashed lines at 316, 324, and 332 K) and divergence (filled).</td>
<td>H. F Dacre, M. K. Hawcroft, M. A. Stringer, and K. I. Hodges. An extratropical cyclone atlas: A tool for illustrating cyclone structure and evolution characteristics. Bull. Amer. Meteor. Soc., 93:1497–1502, 2012.</td>
</tr>
<tr>
<td>53</td>
<td>figure</td>
<td>FIG. 5. Total EKE vs MAPE for all simulations. The series of simulations in which the longwave optical thickness is varied is shown with circles, and that in which the meridional insolation gradient is varied with asterisks. The solid line shows a linear relation through the origin with slope 2.4 (the rescaling factor used for MAPE in Figs. 3 and 4).</td>
<td>Paul A. O’Gorman and Tapio Schneider. Energy of midlatitude transient eddies in idealized simulations of changed climates. J. Climate, 21:5797–5806, 2008b.</td>
</tr>
<tr>
<td>56</td>
<td>figure</td>
<td>Figure 3. Same as Figure 1 but for extreme events corresponding to (a) positive and (b) negative meridional velocity anomalies at the location of the extreme event local maxima. Unlike in Figure 1, the contour intervals are every 5 m s⁻¹ for meridional velocity anomalies and 3 K for MSE ones.</td>
<td>G. Messori and A. Czaja. On local and zonal pulses of atmospheric heat transport in reanalysis data. Q. J. R. Met. Soc., 2015. doi: 10.1029/2002JD003026.</td>
</tr>
</tbody>
</table>