UNIVERSIDAD NACIONAL DE COLOMBIA

Is the Hypothesis of the Condensation-Induced Atmospheric Dynamics a New Theory of the Origin of the Winds?

Alejandro Jaramillo Moreno

Universidad Nacional de Colombia
Facultad de Minas, Department of Geosciences and Environment
Medellín, Colombia
2017

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Alejandro Jaramillo Moreno

A thesis submitted in partial fulfilment of the requirements for the degree of:
Doctor of Engineering in Water Resources

Director: (Ph.D.) Oscar José Mesa Sánchez

Research Track: Hidroclimatology
Research Group: Posgrado en Aprovechamiento de Recursos Hidráulicos (PARH)

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Medellín, Colombia
2017

"Scientific criticism is justified only by establishing truth in place of error. Generally speaking, new principles emerge from the ruins of those abandoned, based strictly on facts correctly interpreted."

Ramón y Cajal (2004)

"Dubitando ad veritatem pervenimus"

Cicero

Abstract

Makarieva and Gorshkov (2007) proposed a new scientific hypothesis where a "previously unstudied property" associated with condensation is fundamental to a full understanding of atmospheric dynamics, and that appears in a confrontation with the standard theory, where the buoyancy force is the dominant driver of atmospheric motions. In this hypothesis condensation of water vapor in the Earth's gravitational field is the driver of low-level air circulation, explaining phenomena like cyclones, monsoon circulations, and even the Hadley circulation. This hypothesis immediately attracted the attention of the academic community, especially of the hydrologists interested in a better understanding of the interaction between the hydrological cycle and the atmospheric circulation. This hypothesis also has received much criticism mainly from the meteorological expert community, which criticizes the validity of this hypothesis and the existence of this "unstudied property." Although this proposed mechanism is controversial, we could not find any paper nor discussion that clearly shows that this hypothesis is physically wrong. In this thesis, we demonstrate that the proposed force is not new but is not effective because its contribution is canceled out in the buoyancy force. Therefore, this force does not play any role in the atmospheric circulation. We show that the description associated with this force is affected by serious problems in its physical formulation including violation of Newton's third law. We also reiterate that the role of the water cycle in the standard theory is essential to explain major atmospheric circulations, but without physical inconsistencies.

Keywords: Biotic Pump Hypothesis, Condensation, Buoyancy, Thermally Driven Circulations, Atmospheric Moist Thermodynamics

Resumen

Makarieva and Gorshkov (2007) proponen una nueva hipótesis científica donde una "propiedad previamente no estudiada" asociada con la condensación es fundamental para una comprensión completa de la dinámica atmosférica, y que aparece en una confrontación con la teoría estándar, donde la fuerza boyante es el principal conductor de los movimientos atmosféricos. En esta hipótesis, la condensación del vapor de agua en el campo gravitacional de la Tierra es el conductor de la circulación de aire de bajo nivel, explicando fenómenos como ciclones, circulaciones monzonicas e incluso la circulación de Hadley. Esta hipótesis atrajo inmediatamente la atención de la comunidad académica, especialmente de los hidrólogos interesados en una mejor comprensión de la interacción entre el ciclo hidrológico y la circulación atmosférica. Esta hipotesis También ha recibido numerosas críticas principalmente de la comunidad de meteorólogos expertos, que critica la validez de esta hipótesis y la existencia de esta "propiedad no estudiada". Aunque este mecanismo propuesto es controvertido, no pudimos encontrar ningún trabajo ni discusión científica que demuestre claramente que esta hipótesis es físicamente incorrecta. En esta tesis, demostramos que la fuerza propuesta no es efectiva porque su contribución se anula en la fuerza boyante. Por lo tanto, esta fuerza no juega ningún papel en la circulación atmosférica. Mostramos que la descripción asociada con esta fuerza se ve afectada por serios problemas en su formulación física, incluida la violación de la tercera ley de Newton. También reiteramos que el papel del ciclo hidrológico en la teoría estándar es esencial para explicar las principales circulaciones atmosféricas, pero sin inconsistencias físicas.

Palabras clave: Hipotesis de la Bomba Biotica, Condensación, Boyancia, Circulaciones Térmicamente Directas, Termodinámica Atmosférica Húmeda

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Preface

> *"In science it often happens that a scientists say, 'You know that is a really good argument; my position is mistaken', and then they would actually change their minds and you never hear that old view from them again. They really do it. It does not happen as often as it should, because scientists are human and change is sometimes painful. But it happens every day. I cannot recall the last time something like that happened in politics or religion."*
> **Carl Sagan**

A Ph.D. is more than a thesis dissertation; it is a learning path that transforms a graduate student into a new researcher. It comprises numerous different experiences that teach, under the helpful guidance of a wise advisor, the fundamental tools to build successful research. These tools cover from identifying a research problem to the development of ideas and methodologies that will guide the path to find answers to the initial questions. The learning ends with the most important lesson: the researcher aspirant must learn to conclude his or her work and to share the results with the scientific community, usually as a thesis dissertation or as a scientific publication. Most of the time the research question leads to a fruitful path that creates new knowledge and advances science, opening new routes for more questions and more research. Sometimes the trail ends abruptly in a deadlock, but it is part of the researcher's work to conclude something about its work, even if it is a negative result. The ability of the researcher to explain the negative results it is also a contribution to science, it will help other scientists following similar approaches to rapidly realize the unfruitful path they were chasing or the need for a different approach that maybe can provide some answers.

This thesis dissertation is part of the results of research that started when I began my Ph.D. studies at the Universidad Nacional de Colombia and will be accompanied by some publications in specialized journals. Nevertheless, I admit that a thesis dissertation does not do justice to the enormous amount of work and thoughts that led to the final conclusions that are here presented. Sometimes, it is valuable to understand not only the conclusions but the different paths that were followed in the trail to the final conclusion. I consider that this preface is the right place to show you a little bit of this thought and argumentation process that summarized, in a certain way, part of the learning process that comprises a Ph.D.

The Biotic Pump Hypothesis[1]

At the very beginning of my Ph.D., my thesis advisor, Oscar Mesa, presented to me a fascinating research question. Two Russian scientists proposed that all the scientists before them were ignoring a fundamental property associated with water vapor condensation so essential that they could explain all the most significant atmospheric flows. They used this principle to explain what seemed to be a revolutionary idea: forest makes use of this property to suck in moist air from the ocean, regulating its precipitation and therefore its climate. This proposal immediately offered a mechanism to solve the old question about the relationship between forest and atmosphere from a very simple perspective and by using simple physics principles.

It was clear that this proposal was in a direct confrontation with the current knowledge of atmospheric physics. The debate was intense with a group of very respected experts in the field that were openly against these ideas and a group of supporters with high expectations of what this hypothesis has to offer. A question immediately appeared: how is it possible that in such an active research area, there is a space for a fundamental property ignored by so many excellent scientists and for so long? This was certainly a subject where a profound understanding of essential physics and thermodynamics is necessary to find the subtle differences in arguments that led to such different conclusions. I was then powerfully attracted to study these novel ideas, and I decided to work on them as my Ph.D. research project. I started to read all the related works and discussions, following carefully the arguments and even the mathematical calculations. I must confess that these authors are hard to read. I guess that in a rush to present their ideas and their arguments, they gave a chaotic view of their thoughts. The process to understand what they truly wanted to say was slow and indeed painful, but in the end, I found that the arguments seemed, at first, convincing.

[1] In the scientific literature, these ideas are known as the biotic pump theory and the theory of condensation-induced atmospheric dynamics. I will prefer to use the term hypothesis instead of theory. The difference between the terms is subtle but important. Following a definition of the scientific method close to the ideas presented by R. Feynman in his Messenger Lectures at Cornell University in 1964, we can understand a scientific hypothesis as a guess about how nature works. We compute the consequences of the guess, and we compare with nature. If it disagrees with experiment, it's wrong. A theory can be considered as a scientific hypothesis that has survived the test of nature but as pointed by S. Hawking: "Any physical theory is always provisional, in the sense that it is only a hypothesis: you can never prove it. No matter how many times the results of experiments agree with some theory, you can never be sure that the next time the result will not contradict the theory. On the other hand, you can disprove a theory by finding even a single observation that disagrees with the predictions of the theory" (Hawking, 1998).

Looking for ways to test the hypothesis

Following the spirit of the scientific method, the next step was to test the hypothesis. Oscar and I started our research work by exploring possible ways to test the hypothesis. Our first ideas were to test by using observations, but we rapidly reached the conclusion that both the classical approach and the Biotic Pump hypothesis would offer similar explanations for the observations, and therefore, this method would not provide a convincing proof for these ideas.

Our next approach was to test the hypothesis by studying a meteorological phenomenon where condensation is a fundamental process and where the current understanding still has problems to explain it. We considered that if the Biotic Pump hypothesis could succeed where the classical approaches had failed, it would constitute a meaningful proof that the ideas behind the hypothesis are necessary and that the principle associated with condensation is worth further exploration.

We found that the Madden-Julian Oscillation (MJO) was a meteorological phenomenon that has some attractive characteristics. The MJO is the dominant component of the intraseasonal (30−90 days) variability in the tropical atmosphere and consists of large-scale coupled patterns in atmospheric circulation and deep convection (Zhang, 2005). This phenomenon has been studied extensively, mainly by means of observations and statistical analyses (e.g., Madden and Julian, 1971, 1972; Hendon and Salby, 1994; Hendon and Liebmann, 1994; Wheeler and Kiladis, 1999; Roundy and Frank, 2004; Kiladis et al., 2005), as well as by the study of individual MJO events (e.g., Lin and Johnson, 1996; Houze et al., 2000; Yanai et al., 2000).

The observed fundamental characteristics of the MJO can be classified as follows:

- **Active and passive phase:** A system of high convection and precipitation ("Active Phase") moving eastward starting in the Indian Ocean, flanked by regions of weak convection both east and west ("Inactive Phase"). Zonal circulations connect the two phases. Close to the surface, there are anomalous winds to west and east of the large-scale convective system (e.g., Madden and Julian, 1972; Zhang, 2005).

- **Propagation:** A propagation towards the east at an average speed of 5 m/s (Majda and Stechmann, 2012), differentiates the MJO from other phenomena that propagate in this direction as the Kelvin waves, which propagate at speeds of 15 to 17 m/s (Wheeler and Kiladis, 1999).

- **Wind-convection coupling:** The relative phase between large-scale surface winds and the convective center varies during the MJO life cycle (e.g., Knutson and Weickmann, 1987; Rui and Wang, 1990; Hendon and Salby, 1994; Sperber, 2003). When the MJO is in the Indian

Ocean, the convective center is usually located between the west and east winds. When the MJO moves to the Pacific, the West winds tend to prevail at the convective center (e.g., Zhang, 2005).

- **Extension:** The zonal extension of an MJO event covers about 12,000 to 20,000 km (e.g., Rui and Wang, 1990). Only one fully developed MJO event exists in the tropics at any given instant. Occasionally, two weak MJO convective centers can coexist, one starting in the Indian Ocean and the other decaying in the central Pacific (e.g., Wheeler and Hendon, 2004).

- **Period:** The dominant period of the MJO extends in the range between 30 and 100 days. This range reflects that although it is called an oscillation, MJO events do not occur at regular intervals, being highly episodic or discrete events (e.g., Hendon and Salby, 1994; Zhang, 2005).

- **Multi-Scale Structure:** The active phase of the MJO can be seen as a large-scale assembly of small convective systems, moving in all directions. The apparent eastward spread of the system is due to the continuous formation of new convective systems, each more to the east than the previous one (e.g., Madden and Julian, 2012).

- **Geographic preference:** The MJO convective signal is confined between the Indian and Western Pacific oceans because convective instability can only be maintained on waters with high surface temperature. When this temperature is greater than 28°C, the energy source required to maintain convection is guaranteed. This geographical region is characterized by strong convective anomalies sustained for periods of 20 to 30 days (e.g., Hoyos and Webster, 2006).

- **Seasonal cycle:** The MJO shows a clear seasonal cycle. The trajectory of the convective center is located in the respective summer hemisphere. The trajectory during the boreal winter presents less variability and is parallel to the equator, whereas in the boreal summer tends to propagate towards the northeast (e.g., Hoyos and Webster, 2006).

- **Intra-annual variability:** The intra-annual variability of the MJO is highly related to the El Niño Southern Oscillation (ENSO) phenomenon. During the El Niño phase, the activity of the MJO spreads more to the east. Before an El Niño event, MJO activity appears to be more vigorous, whereas it appears abnormally weak after a La Niña event (e.g., Zhang, 2005).

Different mechanisms to explain the MJO have been considered, organized mainly in two schools of thought. One considers the MJO to be an atmospheric response to existing independent forces, such as intraseasonal fluctuations in precipitation associated with the Asian monsoon (Yasunari, 1979), stochastic heat sources (Salby and Garcia, 1987) and intraseasonal perturbations of the extratropics (Lau and Peng, 1987; Hsu et al., 1990). The other school thinks that the MJO creates its own energy source through atmospheric instability, such as Wave-CISK interaction (e.g., Lindzen, 1974; Lau and Peng, 1987) or the interaction between evaporation and surface wind, also known as WISHE (e.g., Emanuel, 1987; Neelin et al., 1987; Wang, 1988).

Despite the efforts made, the MJO presents today great difficulties for its correct representation (Hoyos and Webster, 2006; Zhang, 2005) and there is not yet a theory that satisfactorily explains its properties. The observed characteristics of the MJO and the importance of condensation in this phenomenon offered us a possibility to use it to test the Biotic Pump hypothesis. It is possible, in a first attempt, to speculate on the contributions that would make the condensation-induced dynamics in some of the general characteristics as shown below:

- **Active and passive phase:** As a system of high convection and precipitation, the active phase becomes a center of low pressure, generating circulation with patterns of anomalous winds from both the east and the west. The condensation-induced dynamics can be used to understand these pressure gradients created by the transformation of the power associated with condensation into kinetic energy of horizontal motions.

- **Propagation:** Condensation-induced dynamics can provide new insights into the study of tropical waves, where it is possible that it can explain the generation of tropical waves with propagation speeds smaller than those of Kelvin waves.

- **Period:** The period of the MJO is linked to the spatiotemporal evolution of the surface characteristics of the ocean, which favor the formation of centers of greater evaporation. These will be the source of potential energy that will generate the dynamics of the MJO controlling its period.

- **Multi-Scale Structure:** The condensation-induced dynamics can study the different scales in which the MJO acts, explaining the formation and dissipation of the convective systems, as well as studying the interaction of the large-scale system with the small scale.

- **Geographical preference:** The MJO convective signal is confined between the Indian and Western Pacific oceans, in areas with high sea surface temperatures. These zones ensure the flow of water vapor necessary to maintain the rate of condensation. When leaving this zone, the potential energy source, given by the steam, is weakened and therefore can explain the decay of the MJO as it moves east.

- **Extension:** The zonal extension of an MJO event can be explained by studying the intensity of condensation in the phenomenon and the evaporation gradients in the area. The hypothesis of condensation-induced dynamics can calculate the order of magnitude of this extension.

- **Wind-convection coupling**: Due to the winds generated by the gradients due to condensation, the condensation-induced dynamics can explain the relative phase between the large-scale surface-area winds and the convective center.

- **Seasonal cycle:** The clear annual cycle of the MJO can be understood by looking at the evaporation gradients. In the northern winter, the greatest evaporation is concentrated in the ocean, where the vapor flows are more homogeneous, explaining the MJO lower variability and its movement parallel to the equator. During the summer, increasing evaporation in the Asian continent creates atmospheric circulations that interact with the MJO, increasing its variability and generating propagation towards the northeast.

- **Intra-annual variability:** During the El Niño phase the Pacific warm pool is displaced, relocating the potential energy source needed to generate the condensation-induced dynamics. These changes in the Pacific surface waters can be seen as a space-time change in the source of water vapor and in relocation of sources of potential energy to generate atmospheric dynamics.

We pursued this approach with the expectations of finding the positive result that the biotic pump theory will offer a new tool to study the atmospheric dynamics, but we ended up very soon in a deadlock. The physics formulation of the biotic pump hypothesis is still descriptive and does not offer a real tool to study the dynamics needed to explore a complex problem as the MJO. To overcome this issue, a formulation of the dynamics of the hypothesis was necessary. To construct such dynamics, a complete understanding of the subtleties of the arguments that marked the differences with the traditional approach is fundamental. We found that in the literature, this point was not studied strongly enough, and there was a lot of confusion about the correctness of the theoretical formulation.

At this point, we realized that we could not test a scientific hypothesis if there were serious doubts about its physical formulation of it, and therefore, we focused on studying the dynamic foundation of the hypothesis but now with a more skeptical look. After more than a year of weekly debates between Oscar and me in which we explored different aspects of the hypothesis, we found that there was a subtle difference in the physics that passed undetected by the various experts that participated in the debates. This difference pointed to a fundamental confusion with the process of moist convection and the phenomenological consequences of it in the meteorological fields like pressure. As the opening quote by C. Sagan says, we understood that we were blinded by the hope of a positive result for our test of the hypothesis, and now, confronted with the new evidence, we changed our minds.

In this document, I will present this latter approach that offers, in my humble opinion, a very simple conclusion to a difficult problem. To provide a clear answer as simple as possible, I used the language of traditional textbooks on the subject. This approach, although it might seem at first an oversimplification of what is a complicated problem, allows the clearest way to discuss the kind of subtleties that we aim to address.

Acknowledgements

I would like to express, first and foremost, my sincere gratitude to my advisor, Professor Oscar Mesa. I am indebted to him for his continuous support and guidance during my time at the Universidad Nacional de Colombia. He believed in me even at times when I doubted myself. He was always there, supporting my learning process. I thank him for all his patience and all the scientific discussions around cups of coffee that we had week after week. I learned from his love of science, his intuition, and his constant deep questions. I also learned that, even with all the limitations that a researcher has in a country such as Colombia, we are obliged to work harder and to ask ourselves difficult questions, even questions that nobody is willing to answer.

I am also grateful to Dr. George Kiladis. He was my advisor during my one-year visit to the physical science division (PSD) of the Earth System Research Laboratory (ESRL) at NOAA, Boulder. I have no words to express my admiration for him, not only as the great scientist he is but also as the great human being I found in him. I thank him for allowing me to learn from his work and for opening the doors to show me a different way of doing research. I thank him for his constant and sincere concern about my advances in my Ph.D. Thanks to George, I met some great scientists, like David Raymond, David Adams, Olivier Pauluis and Juliana Dias with whom I had interesting scientific discussions that helped me to open my eyes to new perspectives in my research.

I would also like to thank David Raymond. Our conversations during his visits to Medellín were very instructive and helped me a lot to clarify physics concepts. I thank him for his constant and sincere concern about my advances in my Ph.D. His comments pointing to the violation of Newton's third law by the biotic pump hypothesis were fundamental for the reformulation of the ideas presented in this thesis and led to its final form presented here. I thank David Adams, too, for his kindness and feedback during my visit to the UNAM. I thank him for pointing out the importance of temporal scales in the process associated with convection.

I would also like to thank Anastassia Makarieva. Although she is the primary author of the hypothesis I proved here to be wrong, she was open to discussing her ideas and offered us useful critique about our arguments that helped me to enhance the present work.

I am grateful to my thesis examiners Professors German Poveda, Paola Arias, and David Raymond, who enriched this work through their comments on a previous version of this manuscript and during the thesis defense.

I am grateful to the Universidad Nacional de Colombia and Sapiencia-Enlaza Mundos, which supported my work through partial funding during my Ph.D.

I thank my fellow graduates and friends at the National University of Colombia, especially Victor Peñaranda, Julian David Rojo, Juan José Guerrero and Ricardo Roman. I am also grateful to the department professors and administrative personnel, especially Luis Fernando Carvajal, Germán Poveda, Carlos Hoyos and our department librarian, Diana N. Ospina. The interesting academic and nonacademic discussions with them made my education and life, in general, a lot richer.

I also thank all the friendly people that I met during my visit to the USA, particularly Mary Kiladis, Thomas Morris, Juliana Dias, Maria Gehne, John Albers, Barbara Herrli, Alex McColl and Rita Lombardi. Thanks to them, my experience abroad was a lot richer.

Last but not least, I would like to thank my parents, Gustavo and Myriam, my sister, Liliana, and my girlfriend, Christian, for all their support and love. This work would not have been possible without them.

Chapter 1.

Introduction

A. Makarieva and V. Gorshkov, two Russian scientists of the Petersburg Nuclear Physics Institute, proposed a new scientific hypothesis where a "previously unstudied property" associated with condensation is fundamental to a full understanding of atmospheric dynamics (we will refer to this hypothesis as the Makarieva-Gorshkov hypothesis (M-G) hereafter). In this hypothesis condensation of water vapor in the Earth's gravitational field is the atmospheric driver of low-level air circulation, explaining phenomena like the monsoon circulation and even the Hadley circulation (e.g., Makarieva and Gorshkov, 2007, 2009a,c, 2010; Gorshkov et al., 2012; Makarieva et al., 2013b, 2014). The main motivation for these authors to propose this hypothesis is the well known research question of how forest interacts with climate. Although there has been some progress on those topics, today there is no clear and simple explanation of how is this interaction works, and a physical mechanism is needed.

Makarieva and Gorshkov (2007) explain that this property associated with condensation describes an atmospheric process that resembles the physical principle of a pump, where the energy spent on evaporation supports the moisture pump. They propose that forests through their historical evolution use this principle to suck in moist air from the oceans into the continents, maintaining the hydrological cycle inland, this provides a physical mechanism to explain the relationship between forest and climate. This proposal created great interest and expectations in the academic community, especially hydrologists interested in a better understanding of the interaction between the hydrological cycle and the atmospheric circulation, and the role of tropical forests like the Amazon and the important questions of the global effects of deforestation in climate.

This proposal of a new scientific hypothesis based on a supposedly unstudied property associated with condensation immediately created substantial controversy. The current knowledge of atmospheric dynamics uses well-established physics principles, and is a very active research area with thousands of experts working on questions related to different aspects of atmospheric interactions. The logical question that follows is: how is it possible that in such an active research area there is space for a fundamental property ignored by so many and for so long? To try to answer this question, we have to enumerate first what are the possible explanations for this controversy:

- This proposed property simply does not exist. There is something fundamentally wrong with the proposal.
- This proposed property exists but its effect is unimportant to explain the atmospheric flow.
- This proposed property exists and is fundamental to understand atmospheric dynamics. An essential principle is missing from the current understanding of atmospheric physics.

The current debate moves around these possibilities. We can get an idea of the different arguments for or against by reading the discussion page of Makarieva et al. (2013b)[1]. The editor clearly stated the controversy in a final comment published with that paper that says:

> *"The authors have presented an entirely new view of what may be driving dynamics in the atmosphere. This new theory has been subject to considerable criticism which any reader can see in the public review and interactive discussion [...] Normally, the negative reviewer comments would not lead to final acceptance and publication of a manuscript in ACP. After extensive deliberation however, the editor concluded that the revised manuscript still should be published – despite the strong criticism from the esteemed reviewers – to promote continuation of the scientific dialogue on the controversial theory. This is not an endorsement or confirmation of the theory, but rather a call for further development of the arguments presented in the paper that shall lead to conclusive disproof or validation by the scientific community [...] (1) The paper is highly controversial, proposing a fundamentally new view that seems to be in contradiction to common textbook knowledge. (2) The majority of reviewers and experts in the field seem to disagree, whereas some colleagues provide support, and the handling editor (and the executive committee) are not convinced that the new view presented in the controversial paper is wrong. (3) The handling editor (and the executive committee) concluded to allow final publication of the manuscript in ACP, in order to facilitate further development of the presented arguments, which may lead to disproof or validation by the scientific community."*

An editorial comment of this kind and the acceptance of this paper for publication despite the negative reviews is hardly ever seen in journals of high reputation. This unusual situation clearly indicates us that this is a hard problem, where simple and subtle differences in arguments can lead

[1]See `http://www.atmos-chem-phys.net/13/1039/2013/acp-13-1039-2013-discussion.html`

to such opposing conclusions. At first, it seems that the two first possibilities where the property does not exist or it is unimportant, appear to be the most probable answers to this discussion. The last possibility where the property is fundamental seems at first hard to believe, but in science, the scientific method tells us that we must test the hypothesis before we accept or reject them.

It is important to note that although the negative criticism by experts strongly supports the idea that there is something fundamentally wrong with the physics of this hypothesis, it is hard to find an explicit discussion based on physical principles that clearly point on where is the flaw in the arguments. Perhaps the only work in this direction is Meesters et al. (2009), which analyzes the physical impossibility of the new force proposed by Makarieva and Gorshkov (2007). Nevertheless, Makarieva and Gorshkov (2009c) suggest that Meesters et al. (2009) misunderstood the mechanism raised by MGH. Later publications, like Makarieva et al. (2013b), increased the controversy about these ideas. Consequently, there is a need for a strong physical argument resolving the questions concerning MGH.

The aim of this thesis is to provide a clear discussion to try to solve this debate by using basic and well-established physical principles. We will demonstrate from the entropy equation for moist air that the alleged new force exists, but its contribution is canceled in the net buoyancy force. Therefore, the dynamic effects attributed to this force are not physically possible, pointing to deep problems in the physics of the proposal. We will show that the proposal of the new force by this hypothesis violates several physical principles like Newton's third law. To explain the deficiencies in the physics of MGH, we will restrict our arguments to a simplified view of moist convection following an approach similar to that in standard textbooks in atmospheric thermodynamics, but our approach will be rigorous.

This thesis is organized as follows. Chapter 2 reviews the ideas that motivated M-G, and we will explain the main idea behind the proposed property associated with condensation. Chapter 3 explores the process of adiabatic/pseudo-adiabatic cooling, also known in the literature as adiabatic/pseudo-adiabatic expansion, that is by far the most important process by which the Earth atmosphere condenses water vapor (Emanuel, 1994). This chapter will help us to establish an argument and notation according to standard "common textbook knowledge" and offers an interesting discussion about the importance of condensation. Chapter 4 presents the main arguments about the physical flaws of M-G. In this chapter, we explain the existence of this force and the cancellation of its effects in the buoyancy force and we present the main problems with the physics of this force as presented by the hypothesis. Finally, Chapter 5 presents the main conclusions of this work.

Chapter 2.

M-G Hypothesis

Makarieva and Gorshkov (2007) propose that a “previously unstudied property” associated with condensation of water vapor in the gravitational field of the Earth leads to low-level air circulation from areas of weak evaporation toward regions where evaporation is intense. In this and works that followed the original authors and collaborators constructed a consistent physical picture where this unstudied property drives the atmospheric circulations described as the condensation-induced atmospheric dynamics (e.g., Makarieva et al., 2013b, 2014; Makarieva and Gorshkov, 2009a; Gorshkov et al., 2012; Makarieva and Gorshkov, 2009c, 2010). They also propose that this idea applies to high precipitation systems like cyclones (e.g., Makarieva and Gorshkov, 2009b; Makarieva et al., 2015, 2017), where the lack of this valuable property leads to an incomplete understanding of these systems.

The fundamental ideas of this hypothesis resemble the physical principle of a pump where the energy spent on evaporation supports moisture pump. These ideas were first applied to study the relationship between the forest precipitation and the atmospheric circulation, where the natural forests suck in moist air from the oceans into the continents, maintaining the hydrological cycle. Hence it is also known as the Biotic Pump hypothesis (e.g., Makarieva and Gorshkov, 2007, 2010; Sheil and Murdiyarso, 2009; Makarieva et al., 2013a). The ideas of Condensation-Induced Atmospheric Dynamics and its application to the forest as the Biotic Pump hypothesis have attracted, slowly but increasingly, the attention of the academic community. This idea is of especial interest for hydrologists interested in a better understanding of the interaction between the hydrological cycle and the atmospheric circulation, and what is the role of tropical forests like the Amazon in the global circulation. This chapter briefly reviews the main motivations for these ideas but focuses only on the physical principle that drives the dynamics described by this hypothesis.

2.1. The Missing Driver of the Hydrological Cycle on Land

How moisture is transported from oceans to land, and the precise role of vegetation in this process, constitute key questions in our understanding of the hydrological cycle (Meesters et al., 2009). Makarieva and Gorshkov (2007), following a similar approach like Budyko (1974), Savenije (1995b) and Savenije (1995a) studied how the moisture transported by fluxes of atmospheric air from ocean to land weakens as they propagate inland. They considered that the change of horizontal moisture flux as it moves inland should be proportional to the flux itself and this rate of change corresponds to water that returns by gravity as runoff. Runoff (R) is then related with precipitation (P) and evaporation (E) by a simple moisture balance $P = E + R$. From these considerations, they found that precipitation must decrease exponentially, decreasing in direction from the ocean source to the interior as

$$P(x) = P(0)exp\left(-\frac{x}{\ell}\right), \tag{2-1}$$

where $P(0)$ is the precipitation in the initial point $x = 0$, and ℓ is a scale length that reflects the intensity of precipitation processes. This exponential decay of precipitation describes how a parcel of moist air formed in a region where there is a source for atmospheric moisture like the ocean (Gimeno et al., 2010, 2012, 2013) moves inland transported by air flows, disconnected from the original moisture source and depleting its water content as it precipitates inland.

They also studied vast terrestrial regions covered by natural forest (green transects in Figure **2-1**) like the Amazon basin, the Congo basin and the Yenisey basin, regions that represent the largest remnants of Earth's natural forest cover (Bryant et al., 1997; Makarieva and Gorshkov, 2007). For these transects extensively covered by natural forest, these seems to be no dependence of precipitation P with distance x from the coast, and therefore precipitation does not conform to an exponential decay law for these regions. Similar works showed that this precipitation over forest during the periods of high forest activity is always greater than over the adjacent ocean and that precipitation is constant or slightly increases over forested regions in precipitation transects from the coast to the interior (e.g. Makarieva et al., 2009, 2013a). Although this is an interesting argument about the importance of the relationship between forest and precipitation, the main difficulty of this transect approach is that it constitutes a strong generalization. It ignores variations in landform, and cover types within each transect and some of the transects are against the large-scale air flow ignoring the influence of atmospheric circulation patterns (Meesters et al., 2009; Sheil and Murdiyarso, 2009; Angelini et al., 2011). Regarding this, (Poveda et al., 2014) studied the distribution of average seasonal precipitation from The Tropical Rainfall Measuring Mission Multisatellite Precipitation Analysis (Huffman et al., 2007) along the seasonal dynamics of the CHOCO low-level jet (e.g., Poveda and Mesa, 1999, 2000; Poveda et al., 2005), the Caribbean low-level jet (e.g., Poveda and Mesa, 1999; Amador, 2008), and aerial rivers acting on tropical and subtropical South America (e.g., Marengo et al., 2004). They found that the precipitation is

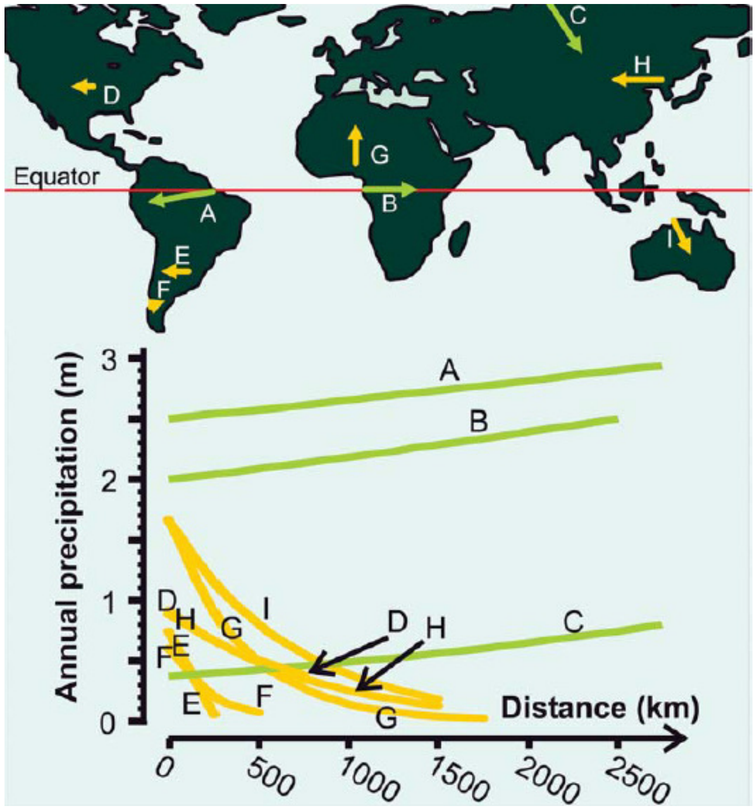

Figure **2-1**.: How precipitation varies with increasing distance inland in forested transects (green) and nonforested regions (yellow). [Permission to use figure from Sheil and Murdiyarso (2009) ©Oxford University Press.]

constant or slightly increases over forested regions and deviations from this behavior can be explained by the effects of topography and land cover types different from forests, supporting the initial argument of Makarieva and Gorshkov (2007) that precipitation is constant over forested regions.

These results reveal a significant difference between the precipitation–distance relationships in the non-forested versus forested regions, where precipitation does not decline with distance from the ocean (Makarieva and Gorshkov, 2007; Makarieva et al., 2009). The law of mass conservation applied to a river basin relates the change of soil moisture content W with time, dW/dt , with precipitation, evaporation and runoff as $dW/dt = P - E - R$ (Manabe, 1969; Poveda and Mesa, 1995; Poveda et al., 2007). In the stationary state $dW/dt = 0$ and runoff R is a function of soil moisture content W, where a high amount of soil moisture implies significant runoff. In areas where soil moisture content does not depend on distance from the ocean, the runoff R is also spatially uniform. If there is sufficient soil moisture, then total evaporation from the dense forest is constrained by solar radiation only, and evaporation should also not depend on the distance from the ocean. Precipitation coupled with evaporation and runoff by the mass balance $P = E + R$ should similarly be independent of the distance from the ocean, a result that is incompatible with an exponential decline of $P(x)$ (Makarieva and Gorshkov, 2007). High precipitation creates high soil moisture content, and forest should be able to transport moisture inland in quantities sufficient to compensate the high runoff losses associated with the maintenance of high soil moisture content (Makarieva et al., 2009). M-G proposes that this is clear evidence that there must be a mechanism used by forest to pump atmospheric moisture inland and therefore there is a feedback between forest and atmospheric circulation that allows the forest to regulate its climate.

This idea of climate regulation by the forests proposed by M-G where forest and atmospheric circulation are strongly coupled is not a new idea. The hydrological and meteorological role of forests has attracted considerable attention from the public over the last two centuries (Andréassian, 2004; Angelini et al., 2011). Andréassian (2004) made an interesting analysis of the historical evolution of ideas, from the ideas discussed in the first century AD by Pliny the Elder to the debates in the France of the 19th century. Numerous other figures throughout history like Christopher Columbus, Noah Webster, Thomas Jefferson, and Alexander von Humbolt have also commented on possible climate change resulting from deforestation (Dickinson, 1989; Angelini et al., 2011). Today the existence of this role played by vegetation in influencing climate is not disputed, constituting a field of active research for a profound physical understanding of its connections and implications (Hayden, 1998; Hutjes et al., 1998; Friedlingstein et al., 2003; Cox et al., 2004; Schneider, 2004; Feddema et al., 2005; Pielke, 2005; Matthews, 2006; Marengo, 2006; Bonan, 2008; Salazar, 2011).

We can classify the different approaches to analyze this relationship between forests and climate in three classes:

- **The observational approach:** Observation and data support these ideas about the relationship between forest and climate. Examples of this kind of study are the mentioned transect analysis in forested regions (e.g., Makarieva and Gorshkov, 2007; Makarieva et al., 2009; Poveda et al., 2014). Poveda and Mesa (1997) suggest that meteorological variables imply the possibility for a mechanism by which the surface processes and the hydrology of tropical South America play a major role in the atmospheric circulation by constructing the "land–atmosphere bridge" connecting sea surface temperatures in the Pacific to those in the tropical North Atlantic and Caribbean. Another example worth mentioning is the work of Worden et al. (2007), that through the study the variation of isotopic content in water vapor over the continents, found that in regions like the Amazon and tropical Africa continental vapor observations are typically less depleted of Semiheavy water(HDO) than the oceanic observations. This observation is contrary to what is expected for an air mass as it moves inland disconnecting from its oceanic source, becoming more depleted. They argue that two possible sources to explain this enrichment of isotopic composition are oceanic vapor transported at low altitude inland and vapor from evapotranspiration.

- **The use of numerical models:** Numerical models attempt to represent the many processes that produce climate as a series of equations that describe the basic physical, chemical and biological principles that define the system (Kendal and Henderson-Sellers, 2005). Pioneering works that studied the relationship between vegetation and climate using numerical models focused on the sensitivity to changes in the surface hydrological regime and energy partition (e.g., Otterman, 1974; Charney, 1975; Shukla and Mintz, 1982; Henderson-Sellers et al., 1988). Works like Charney (1975) and Otterman (1974) suggest that a decline in the surface density of vegetation will alter the surface albedo, enhancing the transfer of radiation back to space and reducing clouds and rainfall. The alteration of the surface water budget induces changes in the large-scale circulation that feedback in a positive sense by causing further vegetation declines through reduced precipitation (Sagan et al., 1979; Gash and Shuttleworth, 1991; Mesa, 2016). Subsequent works have focused on improving the representation of the interaction between biota and dynamics in the equations and parameterizations of the models, where the models are constructed to represent the climate with vegetation present or absent. Numerical experiments are changing the traditional thought that only local climate factors determine vegetation by showing that the presence of vegetation can influence regional climate (Shukla et al., 1990; Nobre et al., 1991). For example, the simulations by Koster and Suarez (1995) suggest that surface processes contribute significantly to the variance of annual precipitation over continents and most studies performed so far imply that extensive deforestation will result in a local reduction in precipitation (e.g., Henderson-Sellers and Gornitz, 1984; Dickinson and Henderson-Sellers, 1988; Nobre et al., 1991; Gash and Shuttleworth, 1991; Xue et al., 2006).

- **The proposal of a theoretical mechanism:** Theoretical proposals are an important way to simplify what otherwise would be a very complex problem, by isolating the main physical ideas to studying the connection between forest and climate. Perhaps the most representative theories in this debate are the works by M. Budyko on the relation between life and the climate (Budyko, 1974), the theory of Gaia of J. Lovelock (Lovelock, 1979; Schneider, 2004) and the Biotic Pump Hypothesis by A. Makarieva and V. Gorshkov that we are discussing in this thesis (Salazar, 2011; Mesa, 2016).

It is clear that Observations and Data suggest that there is a connection between biota and climate, but they do not offer an explanation of what are the causes of these observations. Numerical models on the other hand, if they have a correct representation of the physics, can solve the complex interactions that relate biota to the climate. The main issue is that the models do not numerically resolve the scales on which these interactions act, and they have to use parameterizations of these scales that make difficult or almost impossible the task to identify the mechanisms that explain the interactions between biota and climate. Theoretical proposals try to simplify the physical mechanisms to have a clear understanding of these complex connections between biota and climate. In particular, M-G proposes that a previously unconsidered property of condensation produces a vertical force that they called the evaporative force (sometimes also known as the condensation force). They propose that this force is the missing physical mechanism that can explain how the forest pump moisture from the ocean to the continent, explaining the existence of forest like the Amazon, which with an exponential decay of precipitation would otherwise not exist. They also point out that current atmospheric models ignore this force and therefore they are not suitable to study the connections between forest and atmospheric dynamics. The next section briefly explains the main ideas about this evaporative force.

2.2. The M-G Physical Principle: The Evaporative Force

M-G is based on alleged "unstudied properties" of atmospheric water vapor, proposing the existence of a new and overlooked "evaporative force", that they propose as the main driver of atmospheric dynamics. This new force, and therefore the real role of condensation in the dynamics, seems to be absent in the current common understanding of global circulation. Makarieva and Gorshkov (2007) and Makarieva et al. (2013b) summarize the main physical arguments of this hypothesis, but there are also ample discussion about this ideas by their authors and collaborators in other publications (e.g, Makarieva et al., 2014; Makarieva and Gorshkov, 2009a; Gorshkov et al., 2012; Makarieva and Gorshkov, 2009c, 2010).

The main physical arguments of M-G center on the concept hydrostatic balance in an atmospheric air column. The works related to M-G describe two kinds of balances in a column of static air: The hydrostatic balance and a second kind of balance that M-G defines as "aerostatic balance". The hydrostatic balance is simply the balance between the vertical gradient of the total gas pressure and the weight of the mixture. The aerostatic balance instead, is the balance of the vertical pressure gradient of each of gas component with the corresponding weight of each gas. In a static column, the fulfillment of the aerostatic balance for all the elements of the gas (i.e. the components are in aerostatic balance) implies the fulfillment of the hydrostatic balance of the air as a whole, but the fulfillment of hydrostatic balance does not imply the aerostatic balance of its components.

For large-scale motions, the Earth's atmosphere is to a high degree of accuracy in hydrostatic equilibrium (Holton and Hakim, 2013). Because the dry component of the lower atmosphere is well mixed below the turbopause, the concentrations of the gases that compose the dry component of the air are relatively constant throughout this layer, and it is possible to assign an apparent molar mass to dry air since dry air as a mixture behaves like an ideal gas (Curry and

Webster, 1999). This constant apparent molar mass makes the pressures and densities of the each of the components of the gas mixture to decrease with altitude at the same rate and with a scale height inversely proportional to the apparent molecular weight of the mixture (Wallace and Hobbs, 2006). Therefore dry air is, as a good approximation, in hydrostatic balance but due to this scale height inversely proportional to the apparent molecular mass of the mixture, the components are not in aerostatic balance. For dry air the closeness of the different molar masses with the molar mass of the mixture makes the dry air components to be only slightly off from aerostatic balance.

Water vapor is a highly variable trace gas that can constitute from 0 to 4% of the atmospheric concentration of gases (Curry and Webster, 1999). Moist air can be approximated as an ideal gas when there is no condensation, and the effect of water vapor is usually introduced as a correction in the temperature, known as the virtual temperature correction. However, when the air is saturated, water vapor easily condenses under typical atmospheric conditions restricting the thermodynamics changes of a moist parcel undergoing condensation. We can use the Clausius-Clapeyron to study how the saturated vapor pressure changes during condensation as a function of temperature. Due to the steep decline of air temperature with height in Earth's atmosphere, condensation prevents the aerostatic equilibrium of water vapor in a saturated moist column. This violation of aerostatic equilibrium is manifested as a strong compression of the vertical distribution of water vapor as compared to the distribution of dry atmospheric air and therefore leading to M-G to propose the appearance of an uncompensated vertical force acting on the air called "the evaporative force". The authors state that this force remains practically undiscussed or absent in the meteorological literature.

This evaporative force is described by as the resultant of the difference between the vertical gradient of water vapor pressure and the weight of the vapor component (See Makarieva and Gorshkov (2007) for details):

$$f_e = -\frac{\partial p_v^*}{\partial z} - \rho_v g, \tag{2-2}$$

where p_v^* is the saturated water vapor partial pressure and ρ_v is the vapor density. Another definition for this force is given by (e.g., Makarieva et al., 2013b; Gorshkov et al., 2012)

$$f_e = \frac{p_v^*}{p}\frac{\partial p}{\partial z} - \frac{\partial p_v^*}{\partial z} = -p\frac{\partial \gamma}{\partial z}, \tag{2-3}$$

where p is the total air pressure and $\gamma = p_v^*/p$.

The existence of the force f_e in a stationary state is only possible by maintaining continuous evaporation from the surface, compensating the condensation of the ascending water vapor molecules. If evaporation fluxes in two adjacent areas are different, the ascending fluxes of moist air are different as well. This vertical movement forces horizontal motion translating the condensation-induced pressure difference to a horizontal pressure gradient and forcing horizontal fluxes of moist air. The resulting directed moisture flow will enhance precipitation in the area with strong evaporation and diminish precipitation in the area with weak evaporation.

In contrast with the evaporative force, M-G affirms that buoyancy force averaged over a horizontal scale exceeding the characteristic height of the atmosphere results in a mean Archimedean force of zero. They explain that the transformation of the potential energy of buoyancy into kinetic energy of horizontal motion is strongly inhibited, and therefore the effect of buoyancy on the circulation of atmospheric air only corresponds to a minor correction compared with the contribution of the evaporative force.

2.3. The Evaporative Force and the Atmospheric Circulation

M-G offers a mechanism to explain the connection between forest and climate by introducing the evaporative force described in the previous section. By using this force and the physical principles that they associate with this force, M-G offers a physical picture to explain that this force is, as they argue, the main driver of atmospheric dynamics. If evaporation fluxes in two adjacent areas are different, then ascending fluxes of moist air are different as well. Moist air will flow from areas of weak evaporation to the region where evaporation is more intense.

Makarieva and Gorshkov (2007) presented some examples of how the evaporative force can be used to explain different atmospheric phenomena like the cases shown in figure **2-2**: **(a)** explains the existence of deserts bordering with the ocean. Desert evaporation is negligible compared with the evaporation of the oceanic surface, and therefore the evaporative force is always greater over the ocean[1]. The air flow depletes the desert moisture maintaining the conditions for the desert state to persist in time. **(b)** and **(c)** represent the case of less arid zones like savannas, where evaporation strongly varies from winter to summer seasons. In winter time **(b)**, due to the differences in thermal inertia, the ocean can be warmer than land, and in such a case the evaporative force will be higher over the ocean resulting in a horizontal flux from the land to ocean. In summer time **(c)**, land surface heats up more quickly than the ocean and the evaporative force will be higher on land, resulting in a horizontal flux from ocean to land enhancing precipitation. This variation between winter and summer describes the basic idea of a monsoon circulation corresponding to a dry season and a rainy season. **(d)** represents the Hadley circulation, where the increase of

[1]However, a hot desert next to a cold ocean has a reverse circulation like the Atacama desert next to the cold Pacific (D. Raymond, Personal communication, 2017)

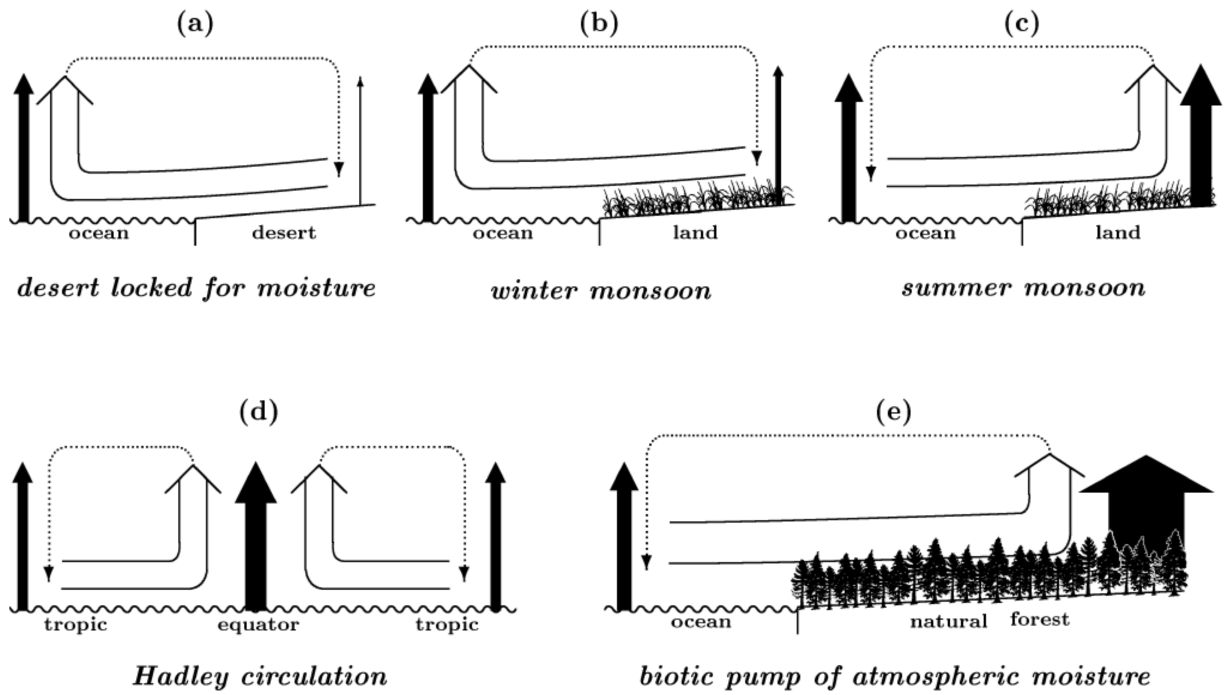

Figure **2-2**.: M-G provides clues for the observed patterns of atmospheric circulation where moist air moves from regions with weak evaporation to areas where evaporation is intense. Some examples are (a) Deserts, (b) Winter monsoon, (c) Summer monsoon, (d) Hadley circulation and (e) Biotic pump of atmospheric moisture. Source: Makarieva and Gorshkov (2007) CC BY-NC-SA 2.5

the solar flux towards the equator increases the evaporative force producing horizontal air fluxes from higher latitudes towards the equator. Finally, **(e)** represents the circulation over a natural forest, which due to the high leaf area index, can evaporate more than an open water surface of the same area. Forest evaporation can be higher than the evaporation in the bordering ocean forcing moist air to flow from the ocean to the forest maintaining the high humidity needed for the existence of the vegetation.

The ideas presented by M-G seem to describe a consistent picture of how this proposed force drives the circulation, giving an explanation that appears to fit with the observed circulation phenomena but by using a new overlooked force. In their review of the ideas of the Biotic Pump hypothesis, Sheil and Murdiyarso (2009) summarized some important implications of M-G that make these ideas relevant for further study. These are:

- **Water yields:** M-G suggests that deforestation will produce a decline in rainfall, also reducing runoff. Sheil and Murdiyarso (2009) suggest that the standard view proposes that deforestation results in less water lost to evaporation, increasing local runoff (Calder, 2005) but works like Poveda and Mesa (1995) explain how deforestation and the loss of water by evapotranspiration are completely indispensable to keep in balance the hydrological cycle supporting the proposal of M-G.

- **Fire:** Fire reduces the ability of the vegetation to maintain humidity reducing evaporation and therefore reducing rainfall, leading to increased droughts, greater flammability, and increased fire risk.

- **Vegetation feedbacks:** M-G implies that changes in the rates of evaporation will influence climate.

- **Evolution and emergent stability:** M-G proposes that the biotic pump mechanism appeared as an evolutionary advantage during the evolution of trees in Earth's history that offers to the forest the ability to generate rich, self-watering terrestrial habitats.

- **Managed vegetation**: M-G propose that only natural and intact forests can maintain a working atmospheric pump, deforestation destroys this mechanism.

- **Greening deserts**: M-G implies that the establishment of a forest will initiate a biotic pump powerful enough to water it.

As alleged by M-G authors, this force seems to be ignored in the conventional view of atmospheric dynamics and opens the controversy about these ideas. An interesting point to note here is that the dynamics that M-G describes resembles what the conventional view of moist convection says about these atmospheric circulations. In the next chapter, we will explore the main basic ideas behind classical moist convection.

Chapter 3.

The Standard Theory of Condensation

In this section, we will study the thermodynamic relationships between changes in pressure, temperature, and condensation for an air parcel of moist saturated air that is condensing under a pseudo-adiabatic process which is by far the most important process by which the Earth's atmosphere condenses water vapor (Emanuel, 1994). Adiabatic expansion occurs when a mass of air (dry or moist) rises where a variety of mechanisms like mechanical lifting (e.g., by topography), turbulent mixing, large-scale dynamic instabilities and buoyancy caused by surface heating (Curry and Webster, 1999; Emanuel, 1994) may cause this rise of air parcels in the Earth's atmosphere. We will study how condensation occurs in an adiabatic expansion, following a classical textbook approach (e.g., textbooks such as Curry and Webster (1999), Dutton (2002), Iribarne and Godson (1981), Wallace and Hobbs (2006) and Emanuel (1994)). From this basic approach, we will show how condensation enters in the dynamics of atmospheric air flow, and we will contrast this with the approach of M-G. Hereafter we will refer to this approach as the standard theory to differentiate it from M-G.

3.1. The Pseudo-Adiabatic Condensation

The specific entropy for moist air above freezing is given by (Raymond, 2013b)

$$\eta = (c_{pd} + r_v c_{pv} + r_L c_L) \ln [T/T_F] - R_d \ln [p_d/p_R] - r_v \ln [p_v/e_{SF}] + \frac{L_L[T_F] r_v}{T_F}, \tag{3-1}$$

where η is the specific entropy per unit of dry air, c_{pd} and c_{pv} are the respective specific heats of dry air and water vapor at constant pressure, c_L is the specific heat of liquid water, r_v and r_L are the mixing ratios of vapor and liquid water, R_d and R_v are the gas constants for dry air and water vapor, $L_L[T_F]$ and e_{SF} are respectively the latent heat of condensation and the saturation vapor pressure at a reference temperature T_F that is usually taken to be the freezing point temperature, and p_R is a reference pressure usually taken to be 1000 hPa.

The ideal gas equation for this mixture is given by :

$$p\alpha = R_d T \delta_c, \tag{3-2}$$

where

$$\delta_c \equiv \frac{1 + r_v/\varepsilon}{1 + r_v + r_L},$$

α is the specific volume of the mixture, δ_r is a term that comes applying Dalton's law to the mixture of dry air and water vapor and condensate water. In the scientific literature δ_c is sometimes introduced as a correction factor to the dry gas constant $R_a = R_d \delta_r$ or more usually as a correction to the temperature, called density temperature $T_v = T\delta_r$ or virtual temperature when $r_L = 0$. A detailed explanation and discussion of how this expression is obtained and its implications for moist and cloud densities comparisons is given in Appendix A.

In the Earth's atmosphere the mixing ratios of water vapor r_v and liquid water r_L are small (r_v is observed to be generally less than 0.04 and r_L is typically of the order $\mathcal{O}(10^{-3})$ (Emanuel, 1994)). These ranges allow us to get a simplified form for the specific entropy neglecting the terms with r_L and r_v everywhere except where multiplied by the latent heat of condensation L_L. The contribution of water vapor partial pressure is small, and we can neglect it, therefore the total air pressure is approximately equal to the dry air pressure and the water vapor mixing ratio r_v is approximately equal to the specific humidity q. We can also ignore the variations of the specific heats at the typical range of Earth's troposphere conditions. In differential form, the simplified version of the moist entropy equation reduces to:

$$d\eta = c_{pd} d(\ln[T]) - R_d d(\ln[p]) + \frac{L_L[T_F]}{T_F} dq. \tag{3-3}$$

Under this approximation the ideal gas law is also is simplified and reduces in differential form to:

$$d(\ln[p]) + d(\ln[\alpha]) = d(\ln[T]), \tag{3-4}$$

where α is the specific volume of the mixture. These forms of the entropy and ideal gas equations are common approximations in most of the standard textbooks in atmospheric thermodynamics (e.g., Curry and Webster (1999), Eq. (6.36); Dutton (2002), p. 276; Iribarne and Godson (1981), p. 84; Emanuel (1994), Eq. (4.7.6)). One common problem of this simplification is that sometimes the entropy equation is used replacing the constant temperature T_F by the actual temperature T in the latent heat term (see last term in eq.(3-3)). This is a common mistake, that implies that the moist entropy and dry entropy are not simultaneously conserved in transformations which change temperature but remain non-condensing (Raymond, 2013a).

The specific humidity q can be expressed in terms of the water vapor partial pressure p_v and the total air pressure p and in differential form this relationship is:

$$d(\ln[q]) = d(\ln[p_v]) - d(\ln[p]), \tag{3-5}$$

These equations hold for non-saturated/saturated conditions, but to study condensation we will focus on the ascending of saturated moist parcels of air. To differentiate a variable from its non-saturated or saturated conditions we will add $*$ as superscript when we refer to a variable in a saturated condition. When moist air is saturated the condensation of water vapor obeys the Clausius-Clapeyron relationship, that relates the changes of water vapor partial pressure at saturation with the changes of parcel's temperature. With the same approximations we have used so far the Clausius-Clapeyron equation can be written as

$$d(\ln[p_v^*]) = \frac{\varepsilon L_L[T_F]}{R_d T} d(\ln[T]), \tag{3-6}$$

where $\varepsilon = R_d/R_v = 0.622$ is the ratio of the gas constant for dry air R_d and water vapor R_v. As we did for the specific heats, we neglect the variation of L_L over the range of atmospheric temperatures (Emanuel, 1994). Hereafter L_L will represent the latent heat of condensation evaluated at the reference temperature T_F.

Eq. (3-3) with $d\eta = 0$ together with (3-6) represents the thermodynamic changes for a moist saturated air parcel that is ascending and condensing adiabatically in the atmosphere. During the ascent, the hydrometeors produced by condensation can remain inside the parcel in a completely reversible adiabatic process. We can also consider that the hydrometeors can immediately precipitate out of the parcel but, due to the small heat, mass and momentum carried out by the condensation products, we can consider that, although irreversible, the resulting process is approximately close to adiabatic, and we describe it as a pseudo-adiabatic process (Wallace and Hobbs, 2006).

All the approximations we have included are common in the analysis of the dynamics of saturated air motion and do not affect the conclusions we will deduce. The analysis presented in M-G also made use of these simplified forms of these equations (cf., Makarieva et al. (2013b) and Gorshkov et al. (2012)).

3.2. Relationships between pressure and temperature

We can study the relationship between changes in pressure and temperature for a pseudo-adiabatic process for moist saturated parcel of air using $d\eta = 0$ in (3-3). This relationship can be written as:

$$d(\ln[T]) = \phi[q^*, T]d(\ln[p]), \tag{3-7}$$

where

$$\phi[q^*, T] \equiv \frac{R_d T_F + L_L q^*}{c_{pd} T_F + \frac{\varepsilon L_L^2 q^*}{R_d T}}, \tag{3-8}$$

and ϕ is defined for an easy comparison with the equivalent Eq. (10) in Makarieva et al. (2013b).

Using (3-4) and (3-7) we can get and expression for the change of specific volume in a pseudo-adiabatic process

$$d(\ln[\alpha]) = (\phi[q^*, T] - 1)d(\ln[p]). \tag{3-9}$$

Additionally we can get and expression for the change of saturated specific humidity in a pseudo-adiabatic process as

$$d(\ln q^*) = \lambda[q^*, T]d(\ln[p]), \tag{3-10}$$

with

$$\lambda[q^*, T] \equiv \frac{\varepsilon L_L}{R_d T}\phi[q^*, T] - 1, \tag{3-11}$$

where $d(\ln q^*) < 0$ means condensation of water vapor inside the air parcel.

In Figure **3-1** and Figure **3-2**, $\phi[q^*, T]$ of (3-8) and $\lambda[q^*, T]$ of (3-11) respectively are calculated for a typical range of temperature and saturated specific humidity in the Earth's atmosphere. From the figures, it is clear that they usually are in the range of $0 < \phi < 0.3$ and $0 < \lambda < 5$ for typical atmospheric conditions. These ranges allow us to conclude that for a pseudo-adiabatic cooling the sign of the change of temperature is equal to the sign of the change of pressure and it is opposite to the sign of the change of specific volume. Also, the sign of the change of saturated specific humidity is equal to that of the change of pressure. These relationships describe that for a pseudo-adiabatic expansion, an air parcel is ascending, decreasing its pressure and its temperature but increasing its specific volume. Condensation of water vapor accompanies such decrease of pressure in a saturated pseudo-adiabatic process. Also, If we consider the case with no water vapor, i.e., $q^* = 0$ in (3-7) and (3-9), we get the classical equations for a adiabatic cooling for a dry (or unsaturated) parcel.

Equations of the form of (3-7) shows the relationship between changes of pressure and temperature where the effects of condensation are implicit inside ϕ. We can explore these relationships further if we explicitly separate the contribution of condensation. From (3-3), for an adiabatic/pseudo-adiabatic process the changes of pressure in term of changes of temperature and saturated specific humidity are:

$$d(\ln p) = \frac{c_{pd}}{R_d} d(\ln[T]) + \frac{q^* L_L}{R_d T_F} d(\ln q^*). \tag{3-12}$$

Eq. (3-12) clearly states that the decrease of pressure during the expansion has two components. With the help of (3-7) and (3-10), we can rewrite (3-12) as:

$$d(\ln p) = \frac{c_{pd}}{R_d} d(\ln[T]) + \frac{q^* L_L}{R_d T} \left(\frac{\varepsilon L_L - c_{pd} T}{R_d T_F + L_L q^*} \right) d(\ln[T]) \tag{3-13}$$

When condensation is absent the second term of the right-hand side clearly vanishes leaving only the first term; therefore this first term can be interpreted as the contribution of the dry portion of the process and the second term as the aiding of condensation to the process. The vertical axis in Figure **3-3** represents the term

$$\frac{Cond}{Dry} = \frac{\frac{q^* L_L}{R_d T} \left(\frac{\varepsilon L_L - c_{pd} T}{R_d T_F + L_L q^*} \right)}{\frac{c_{pd}}{R_d}}, \tag{3-14}$$

that represents the comparison of the contribution of condensation compared with the "dry" contribution as shown in (3-13). Both terms have the same sign that correspond to a decrease in pressure with a drop in temperature. It is clear that the condensation contribution is significant and could even be dominant for high values of saturated specific humidity.

3.3. Pseudo-adiabatic condensation and the buoyancy force

To understand how the previous relationships affect the dynamics of the atmospheric flow, we will concentrate on the physics of a moist saturated parcel that is ascending adiabatically/pseudo-adiabatically in a hydrostatic atmosphere. This problem is similar in fluid mechanics to an ascending body submerged in a fluid at rest. Newton's second law per unit volume applied to this body of density ρ, neglecting the effects of the parcel on the surrounding fluid of density ρ' is (Batchelor, 2000):

$$\rho \frac{dw}{dt} = (\rho' - \rho) g - F_f, \tag{3-15}$$

where w is the vertical component of the parcel velocity. The external forces on the body are: the weight per unit volume of the parcel ρg, the upward directed Archimedes' force $\rho' g$, and net frictional force that the surrounding fluid exerts over the body F_f, and always opposite to the direction of motion. We can identify $f_b = (\rho' - \rho) g$ as the net buoyancy force acting on the parcel, which can be positive or negative. Because the surrounding fluid is in hydrostatic balance we can write f_b as:

$$f_b = -\frac{\partial p_e}{\partial z} - \rho g. \tag{3-16}$$

Where the pressure gradient corresponds to the environment. As we did for the body submerged in a fluid, we will study a moist saturated parcel ascending in mechanical equilibrium with a surrounding hydrostatic atmosphere (i.e., the air parcel ascends and expands instantaneously adjusting its pressure to the imposed pressure gradient of the external hydrostatic atmosphere (Salby, 1996)). For a pseudo-adiabatic process, we will ignore the momentum fluxes due to the hydrometeors precipitating out of the parcel, i.e., we are considering the parcel as a closed system, although it is, of course, an open system due to the precipitation[1]. For an adiabatic/pseudo-adiabatic ascent the pressure change for the parcel, from (3-3) with $d\eta \approx 0$, will be a function of the change of temperature T and saturated specific humidity q^* as:

$$dp = \rho c_{pd} dT + \xi p d\gamma, \tag{3-17}$$

where $\xi \equiv \varepsilon L_L[T_F]/R_d T_F \approx 20$ and we use γ (See (2-2)) to easily compare our results with the equations of M-G.

We can study the dynamics for the fluid flow of a continuum of fluid parcels that are ascending and condensing adiabatic/pseudo-adiabatically. Once the flow reaches a stationary state, the zone of ascending flow will consist of an air column saturated at all levels due to the continuous condensation of the ascending parcels. From (3-17) we can obtain the pressure gradients inside this column. The vertical pressure gradient for the stationary state is

$$\frac{\partial p}{\partial z} = -(\rho c_{pd} \Gamma_p + \xi f_e), \tag{3-18}$$

where $\Gamma_p \equiv -\partial T/\partial z$ is the lapse rate of the ascending air parcel and f_e is the definition of evaporative force given by (2-2). The second term of the right-hand side of (3-18) accounts for the

[1]We will emphasize more about this difference between open and closed parcel in the discussion about fluid circulation in the next chapter.

contribution of condensation to the stationary vertical pressure gradient in this flow of ascending parcels. Because the parcels are ascending in mechanical equilibrium with the surrounding atmosphere, this vertical pressure gradient $\partial p/\partial z \approx \partial p_e/\partial z$, i.e, the vertical pressure gradient is imposed by the environment and therefore

$$\frac{\partial p}{\partial z} = -\rho' g = -(\rho c_{pd}\Gamma_p + \xi f_e). \tag{3-19}$$

From this equation is clear that the Archimedes force f_a is equal to

$$f_a = \rho c_{pd}\Gamma_p + \xi f_e. \tag{3-20}$$

The appearance of ξf_e in the vertical component of the stationary pressure gradient force clearly states that condensation of water vapor, in the ascending column, has a contribution to the upward buoyancy force. This contribution is proportional to the deviation of the partial pressure profile of water vapor from the aerostatic balance that M-G states, but this contribution is in fact inside the "common textbook knowledge" as the addition of condensation to the Archimedes' force, and corresponds to the contribution of latent heat release to the parcel's buoyancy. In the stationary state, condensation in the ascent region is manifest as a strong compression of the vertical profile of water vapor. Figure **3-4** shows this compression, where we can compare the vertical profile of p_v created by the steady flow of parcels ascending in an hydrostatic atmosphere with the aerostatic profile that water vapor partial pressure should have if it were in aerostatic equilibrium. The figure shows that condensation in the ascending column strongly constrains the water vapor pressure profile to be away from aerostatic balance, but this compression is a consequence of the steady continuous ascent of moist saturated parcels and does not correspond to an hydrostatic adjustment of the air column. In the ascent column, the vertical pressure profile will be the hydrostatic profile of the environment due to the ascent of moist parcels in mechanical equilibrium with the surrounding atmosphere, but this does not mean that the column is in hydrostatic balance. Figure **3-5** shows the vertical pressure gradient for a column where we have parcels ascending adiabatically compared with the required vertical pressure gradient for the column to be in hydrostatic balance. The small difference between these profiles shows that the column is out of hydrostatic equilibrium.

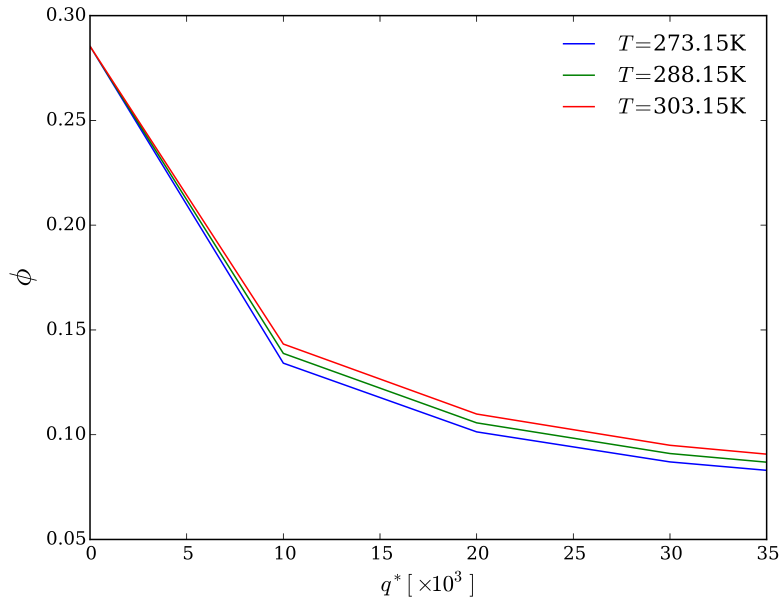

Figure **3-1**.: ϕ calculated for typical range of values of T and q^* in Earth's atmosphere

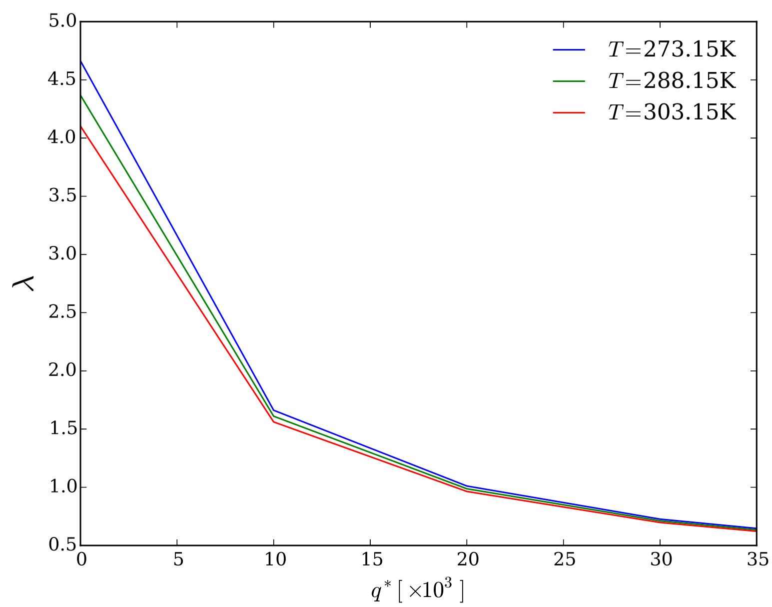

Figure **3-2**.: λ calculated for typical range of values of T and q^* in Earth's atmosphere

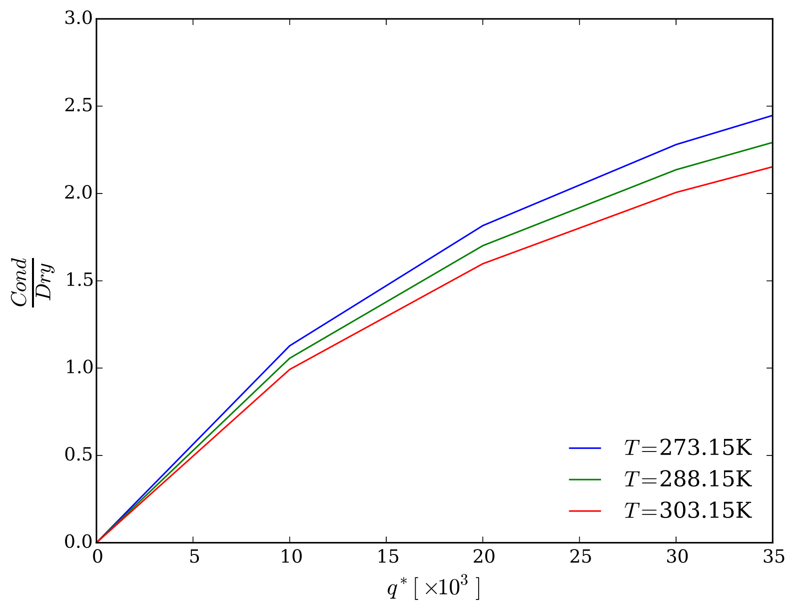

Figure **3-3**.: Comparison from Eq.(3-13) for typical range of values of T and q^* in Earth's atmosphere

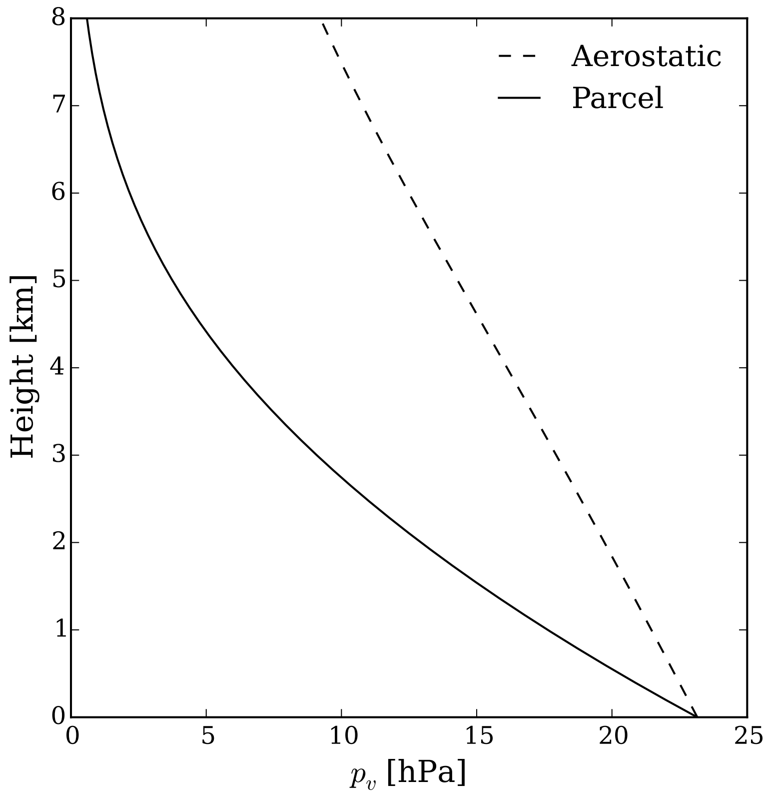

Figure **3-4**.: Vertical aerostatic profile (dashed) of water vapor partial pressure p_v and for saturated moist parcel ascending (solid curves) in an environment with mean tropospheric lapse rate $6.5\,\mathrm{K\,km^{-1}}$ and surface temperature $293\,\mathrm{K}$.

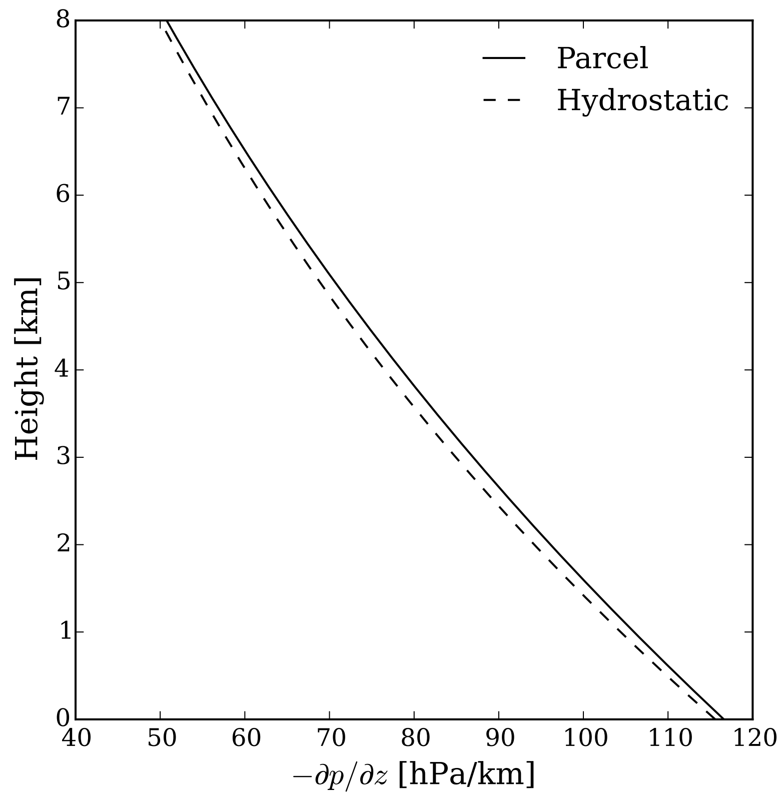

Figure **3-5**.: Comparison between the vertical pressure gradient ($-\partial p/\partial z$) for a moist saturated air parcel in adiabatic ascent (solid curves) and the required vertical pressure gradient for the column to be in hydrostatic balance (dashed). The environment has a mean tropospheric lapse rate $6.5\,\mathrm{K\,km^{-1}}$ and surface temperature $293\,\mathrm{K}$.

Chapter 4.

M-G hypothesis is wrong[1]

In this chapter, we will demonstrate from the entropy equation for moist air that the alleged new force proposed by M-G exists, but its contribution is canceled in the net buoyancy force. Therefore, the dynamic effects attributed to this force are not physically possible, pointing to deep problems in the physics of the proposal. We will show that the proposal of a dynamics based on this force violates Newton's third law. To explain the deficiencies in the physics of M-G, we will restrict our arguments to a simplified view of moist convection following an approach similar to that in standard textbooks in atmospheric thermodynamics as we showed in Chapter 3, but our approach will be rigorous. In this chapter, we also present a review of the main problems with the physics of this force as presented by the hypothesis.

4.1. Revisiting the Buoyancy Force From the Entropy Equation.

As we explained in Chapter 3, the buoyancy force for a parcel of unit mass of moist air is the force due to the difference between the environmental vertical pressure gradient and the weight of the parcel. This may be split into contributions from the dry air and the water vapor as

$$f_b = -\frac{\partial p}{\partial z} - g\rho = -\frac{\partial p_d}{\partial z} - g\rho_d - \frac{\partial p_v}{\partial z} - g\rho_v. \tag{4-1}$$

To understand how condensation enters in this force, we will concentrate on the physics of a moist saturated parcel that is ascending adiabatically in a hydrostatic atmosphere. For a saturated parcel, we replace the vapor pressure p_v with the saturated vapor pressure p_v^* in (4-1) and use

[1] The results of this chapter are presented in Jaramillo et al. (2017) and Jaramillo and Mesa (2017b).

(2-2) to write the buoyancy in terms of the dry quantities and the evaporative force f_e:

$$f_b = -\frac{\partial p_d}{\partial z} - g\rho_d + f_e. \tag{4-2}$$

The model of (Makarieva and Gorshkov, 2007) can be cast as an approximation in which the dry terms in (4-2) are omitted, implying that $f_b = f_e$. We now demonstrate that this is not a good approximation[2].

The specific moist entropy equation links the dry partial pressure p_d, vapor pressure p_v, temperature T and vapor mixing ratio r_v. For saturated moist air, it takes the form (Raymond, 2013b)

$$\eta = (c_{pd} + r_v^* c_{pv}) \ln\left[\frac{T}{T_F}\right] - R_d \ln\left[\frac{p_d}{p_R}\right] - r_v^* R_v \ln\left[\frac{p_v^*}{e_{SF}}\right] + \frac{L_L[T_F] r_v^*}{T_F}, \tag{4-3}$$

where η is the moist entropy per unit mass of dry air, c_{pd} and c_{pv} are the respective specific heats of dry air and vapor at constant pressure, and r_v^* is the mixing ratio of saturated vapor. R_d and R_v are the gas constants for dry air and vapor, L_L and e_{SF} are respectively the latent heat of condensation and the saturation vapor pressure at a reference temperature T_F that is usually taken to be the freezing point temperature, and p_R is a reference pressure usually taken to be 1000 hPa. For simplicity, we ignored the contribution of the liquid water to the specific entropy. Differentiating (4-3) and assuming that the specific heats are relatively constant in the ranges of interest in the Earth's atmosphere, we have

$$d\eta = (c_{pd} + r_v^* c_{pv}) d\ln[T] - R_d d\ln[p_d] - r_v^* R_v d\ln[p_v^*] + \Lambda dr_v^*, \tag{4-4}$$

where Λ collects all terms proportional to dr_v^* and is given by

$$\Lambda = \frac{L_L[T_F]}{T_F} + c_{pv} \ln\left[\frac{T}{T_F}\right] - R_v \ln\left[\frac{p_v^*}{e_{SF}}\right]. \tag{4-5}$$

To study an adiabatic process, we can use (4-4) with $d\eta = 0$. For a column composed of saturated parcels ascending adiabatically in the atmosphere, it is straightforward to show from (4-4) that in the steady state the vertical pressure gradient for dry air inside this column is given by

$$\frac{\partial p_d}{\partial z} = \rho_d (c_{pd} + r_v^* c_{pv}) \frac{\partial T}{\partial z} - \frac{\partial p_v^*}{\partial z} + \frac{\Lambda}{R_d} \frac{\partial r_v^*}{\partial z}. \tag{4-6}$$

In agreement with the parcel model assumption, this is also the vertical pressure gradient in the environment. It may therefore be substituted into (4-2), yielding after some algebra

$$f_b = \left(-\rho_d (c_{pd} + r_v^* c_{pv}) \frac{\partial T}{\partial z} - \frac{\Lambda}{R_d} \frac{\partial r_v^*}{\partial z} - \rho g - f_e\right) + f_e. \tag{4-7}$$

[2] In fact, M-G proposes that f_e is a force independent of the buoyancy force, but it is clear from (4-2) that f_e cannot be seen independent of f_b. M-G is therefore consistent with a proposal where the dry contribution to f_b is omitted, but the arguments proposed by M-G reveals deep problems with fundamental physical concepts.

This expression shows that a component of the force on the dry air exactly cancels the evaporative force f_e that appears on the water vapor. Therefore, the net force on the mixture reduces to

$$f_b = -\rho_d(c_{pd} + r_v^* c_{pv})\frac{\partial T}{\partial z} - \frac{\Lambda}{R_d}\frac{\partial r_v^*}{\partial z} - \rho g. \tag{4-8}$$

It is evident from (4-7) that f_e does not contribute to the buoyancy force due to its exact cancellation. Therefore, the approximation made to the buoyancy equation is a poor one and M-G conflicts with standard atmospheric dynamics and thermodynamics.

4.2. The Inconsistencies of M-G

It is clear from the previous section that M-G is inconsistent with standard atmospheric dynamics and thermodynamics. The necessity of M-G to construct a dynamics based on f_e points to deep conceptual problems that we will pinpoint in this section where we examine M-G at a more fundamental level.

4.2.1. Two Contradicting Definitions for The Evaporative Force

As we pointed out in Chapter 2, M-G proposes two different definitions for f_e. In Makarieva and Gorshkov (2007), M-G proposes the evaporative force as the difference between the water vapor pressure gradient and the weight of water vapor:

$$f_e = -\frac{\partial p_v^*}{\partial z} - \rho_v g = -\frac{\partial p_v^*}{\partial z} - \frac{p_v}{h_v}, \tag{4-9}$$

where in the second equality the water vapor density has been eliminated in favor of the pressure and the scale height of the water vapor.

In later works (e.g., Makarieva et al., 2013b; Gorshkov et al., 2012), M-G uses another definition for this force where

$$f_e = -\frac{\partial p_v^*}{\partial z} - \frac{p_v}{h}, \tag{4-10}$$

where the scale height of the mixture h is used instead of h_v, i.e., M-G is considering that $h_v = h$ but still considering the evaporative force as the difference between the pressure gradient of water vapor and the weight of water vapor. Gorshkov et al. (2012) justifies this equality by saying that “if there is no condensation hydrostatic equilibrium is established in such a manner that at all heights the mean molar mass of the air mixture is equal to its value at the surface, and $h_v = h_d = h$." This justification is physically nonsense and points to a problem of M-G to understand the hydrostatic balance of a mixture of ideal gases.

To explain the problem with the M-G argument we can start with the hydrostatic balance equation which is

$$-\frac{\partial p}{\partial z} = \rho g. \tag{4-11}$$

By using Dalton's law of partial pressures $p = p_d + p_v$ and with $\rho = \rho_d + \rho_v$ we can rewrite (4-11) in two different ways. The first way is

$$-\frac{\partial p_d}{\partial z} - \frac{\partial p_v}{\partial z} = \rho_d g + \rho_v g. \tag{4-12}$$

By using the ideal gas law with $p_i = \rho_i \frac{R}{M_i} T$ where R is the universal gas constant and M_i is the molecular weight of the gas, we can write (4-12) in term of the scale heights $h_i = \frac{RT}{M_i g}$ as

$$-\frac{\partial p_d}{\partial z} - \frac{\partial p_v}{\partial z} = \frac{p_d}{h_d} + \frac{p_v}{h_v}. \tag{4-13}$$

A second way to write (4-11) is by defining a mean scale height for the mixture h that with some straightforward algebra can be showed to be equal to

$$h = \frac{\rho_d h_d + \rho_v h_v}{\rho}. \tag{4-14}$$

By using this mean scale height, we can write (4-11) as

$$-\frac{\partial p}{\partial z} = \frac{p}{h}. \tag{4-15}$$

and using Dalton's law in this equation we get

$$-\frac{\partial p_d}{\partial z} - \frac{\partial p_v}{\partial z} = \frac{p_d}{h} + \frac{p_v}{h}. \tag{4-16}$$

From (4-13) and (4-16) is clear that

$$\frac{p_d}{h} + \frac{p_v}{h} = \frac{p_d}{h_d} + \frac{p_v}{h_v}. \tag{4-17}$$

But from (4-17) it is easy to erroneously assume that $h_v = h_d = h$ and that the following equalities hold:

$$\frac{p_d}{h} = \frac{p_d}{h_d} \text{ and } \frac{p_v}{h} = \frac{p_v}{h_v} \tag{4-18}$$

which are clearly nonsense because h, h_d and h_v are well defined and are clearly different. M-G seems to have followed this erroneous conclusion (e.g. Gorshkov et al., 2012), and therefore writing f_e using h and not h_v as shown in (4-10).

It is clear that the definition of f_e given by (4-10) is not consistent with the original definition where f_e is force as the resultant of the difference between the pressure gradient of water vapor and the weight of water vapor. This equation is basically saying that $p_v/h = \rho_v g$ which is absolutely false and it is straightforward to demonstrate that

$$\frac{p_v}{h} = \frac{h_v}{h}\rho_v g, \tag{4-19}$$

where $h_v > h$ means that $p_v/h > \rho_v g$.

It is important to pinpoint that most of the late works related with M-G use the definition given by (4-10).

4.2.2. The Violation of Newton's Third Law

One of the problematic points of M-G is the role of the dry component that does not seem to be adequately addressed by the hypothesis. For example, in the initial derivation of f_e, Makarieva and Gorshkov (2007) explicitly say that the dry air is in hydrostatic equilibrium (See their Eq. (15)). The work of Meesters et al. (2009) criticized this point because it is not physically possible to talk about a force that disturbs the hydrostatic balances of the air as a whole acting only on the water vapor and leaving the dry air unaffected.

In later works (e.g., Makarieva et al., 2013b; Gorshkov et al., 2012), M-G explains that it is in the absence of hydrostatic adjustment that the dry air is not affected by condensation and remains in equilibrium, but when the hydrostatic balance is restored, the dry air distribution is adjusted at the expense of the horizontal motion. M-G proposes that in the steady state the hydrostatic equilibrium is restored but given by the following equalities (Gorshkov et al., 2012, Eq. (17))

$$-\frac{\partial p}{\partial z} = \frac{p}{h}, \; -\frac{\partial p_d}{\partial z} = \frac{p}{h} - \frac{p_v}{h_c}, \; -\frac{\partial p_v}{\partial z} = \frac{p_v}{h_c}. \tag{4-20}$$

In the real world, the vertical expansion of the water vapor column due to the difference between its actual and aerostatic scale heights is frustrated by the dry atmosphere. Water vapor molecules slowly diffuse through the dry component, a process that is manifested on microscopic scales by collisions between water vapor and dry air molecules. The aggregate result of these collisions is an upward force exerted by the water vapor on the dry air; this is precisely the evaporative force.

M-G fail to point out that by Newton's third law, an equal and opposite downward aggregate force is exerted by the dry air on the water vapor. The upward force of the water vapor on the dry air and the downward force of the dry air on the water vapor exactly cancel, leaving no net force on the atmosphere as a whole. The evaporative force thus has no fluid dynamical consequences, and the whole superstructure of M-G collapses due to a failure at its foundation. This collapse is not due merely to a failure to conform to classical ideas about geophysical fluid dynamics, but to the violation of a bedrock physical principle, Newton's third law.

Curiously, Meesters et al. (2009) came very close to making a decisive argument against M-G. However, they failed to point out the crucial role of Newton's third law in disproving this hypothesis, allowing spurious arguments by Makarieva and Gorshkov (2009c) to muddy the waters.

4.2.3. M-G Condensation is not Adiabatic

Another problematic point of M-G is that it describes a circulation that resembles the one explained by the standard theory of moist convection, making their arguments to appear reasonable in a first approximation. For example, the steady pressure gradients presented in (4-20) coincide with the pressure gradients found in a moist circulation by following the parcel method, where moist saturated parcels rise in the atmosphere in mechanical equilibrium with an hydrostatic atmosphere. In a Eulerian framework, the steady rise of parcels will produce a description similar to the one described by M-G: (1) A column saturated at all heights because at each level there is a saturated parcel that is ascending and condensing in the column. (2) A compressed profile of water vapor as a consequence of the parcels condensing following the Clausius-Clapeyron relationship. (3) An hydrostatic pressure profile of the total pressure due to the mechanical equilibrium of the parcel with the imposed pressure of an external hydrostatic atmosphere.

The similarity of M-G with the standard moist convection can be explained by revisiting the example presented by Makarieva et al. (2013b) in their section 3. In that example, the M-G explains that due to the condensation forced by an imposed temperature lapse rate, the gradient of water vapor pressure does not balance the weight of the vapor component, creating a force f_e that disturbs the hydrostatic balance of the air as a whole. Nevertheless, it is not clear how an adiabatic condensation due to the introduction of a temperature gradient could exist. As is well known, in an adiabatic process there are not heat fluxes through the boundaries of the parcel. Therefore, the only way a parcel can change its temperature adiabatically is through the expansion or compression with the subsequent condensation as described by (4-4).

The only explanation is that the M-G is using (4-4) with $d\eta = 0$ but violating the conditions of an adiabatic process. Instead of the temperature and condensation changes produced by the imposed changes in pressure during the parcel ascent, M-G is describing the pressure and condensation changes generated by an imposed temperature change. However, the latter is forbidden in an adiabatic process. In other words, the mathematical description is the same, but the physics is wrong.

4.2.4. Role of Evaporation and Precipitation in Atmospheric Circulation[3]

M-G was originally proposed to improve our understanding of the airflow and moisture transport over forests. Atmospheric circulations, including those over forests, are readily explained by the equation for the enthalpy change

$$dh = T\,d\eta + \frac{dp}{\rho} \tag{4-21}$$

and the Kelvin circulation theorem

$$\frac{d\Gamma}{dt} = -\oint \frac{dp}{\rho} - \text{friction} = \oint T\,d\eta - \text{friction}. \tag{4-22}$$

(The dh term drops out in this equation because it is the integral of a perfect differential around a closed loop.)

From (4-22) it is clear that a circulation can be maintained against friction only if entropy η is added at a higher temperature than it is subtracted as we move around the circulation loop.

For a dry circulation, entropy changes are driven by sensible heating and cooling. For a moist circulation with precipitation, a more subtle process takes place. Entropy is removed in the descending branch of the circulation by radiative cooling. It is introduced at low levels by surface fluxes of moist entropy, which are produced by a combination of latent and sensible heat fluxes. Since the surface is typically warmer than the free troposphere where radiative cooling is concentrated, the net effect on the circulation is positive.

Note that precipitation, which has to balance the evaporation, plays a secondary role in the entropy budget. The fallout of precipitation decreases the entropy in the free troposphere, but since vapor has higher specific entropy than condensate, the net effect is still an increase in the entropy from the water cycle.

[3] In Appendix B we present a simple model for a moist circulation inspired in similar examples for dry circulations as studied in textbooks.

This argument shows the importance of surface evaporation to the maintenance of atmospheric circulations over forests and oceans. However, unlike in M-G, the evaporation does not have to be co-located with the precipitation. Thus, for example, evaporation in the Tradewind regions feeds the Hadley circulation, and the deep convection in its ascending branch does not depend solely on surface evaporation occurring directly beneath it.

4.2.5. Continuity Equation

Perhaps the most controversial argument of M-G is the proposal that the horizontal pressure gradients that drive the atmospheric circulation can be explained using f_e[4]. By using the continuity equation and the ideal gas law, M-G shows that the horizontal pressure gradients can be related to the dynamic power of f_e, strongly criticizing a buoyancy-driven dynamics. M-G also calculates straightforwardly the dynamic power of the atmospheric circulation coinciding with the estimated order of magnitude, an example that M-G used to support the correctness of its arguments (e.g., Makarieva et al., 2013b). Here we aim to show the problem in the physics of the equation that the M-G proposes to relate the horizontal pressure gradients with the power of their new force.

A parcel of moist air is a mixture of dry air and water vapor that move together. In this section, we will use the continuity equation to obtain restrictions on the horizontal and vertical gradients of pressure, which the motion of this parcel must fulfill. We will consider that the parcel conserves its dry component during its motion, but the vapor component is not conserved, due to condensation and evaporation. For this motion the water vapor mixing ratio $r_v = \rho_v/\rho_d$, the ratio of the density of water vapor ρ_v and dry air ρ_d, obeys the equation (Raymond, 2013b)

$$\frac{dr_v}{dt} = \frac{1}{\rho_d}\nabla \cdot (K\nabla r_v) + \mathcal{E} - \mathcal{P}, \tag{4-23}$$

where $d()/dt$ is the usual material derivative of fluid dynamics, K is the dynamic eddy mixing coefficient and $\mathcal{E}$ and $\mathcal{P}$ are respectively the evaporation and precipitation rates per unit mass of dry air.

Using ideal gas law, it it easy to write the water vapor mixing ratio in terms of the water vapor pressure and the dry air pressure as

$$r_v = \varepsilon \frac{p_v}{p_d}, \tag{4-24}$$

[4]We have already shown that it is not possible to define a dynamics based on f_e. Nevertheless, in this part, we will show that the proposed dynamics by M-G is based on a clear violation of the continuity equation of fluid dynamics.

where $\varepsilon = R_d/R_v = 0.622$ is the ratio of the dry air and water vapor gas constant. Using (4-24) and writing the parcel's velocity $\mathbf{v}$ in a reference frame with the x axis directed along the horizontal velocity u and with w as the vertical velocity, i.e. as $\mathbf{v} = u\mathbf{i} + w\mathbf{k}$, (4-23) can be written for the steady state as

$$u\left(\frac{\partial p_v}{\partial x} - \frac{p_v}{p_d}\frac{\partial p_d}{\partial x}\right) + w\left(\frac{\partial p_v}{\partial z} - \frac{p_v}{p_d}\frac{\partial p_d}{\partial z}\right) = \mathcal{C}, \tag{4-25}$$

where $\mathcal{C} = (\mathcal{E} - \mathcal{P})p_d/\varepsilon$ and herein we will refer to $\mathcal{C}$ as a net condensation rate in units of Pa s^{-1}. For simplicity we ignored the turbulent mixing term proportional to $\nabla \cdot (K\nabla r_T)$.

Eq. (4-25) is the continuity equation together with the state equation for ideal gases and represents the restrictions that the continuity equation imposes on the pressure gradients due to a given condensation rate. We can give to the two terms of the left-hand side of (4-25) a simple interpretation. If we study pure horizontal motion (i.e. $w = 0$), (4-25) reduces to

$$\mathcal{C}_x = u\left(\frac{\partial p_v}{\partial x} - \frac{p_v}{p_d}\frac{\partial p_d}{\partial x}\right), \tag{4-26}$$

where $\mathcal{C}_x$ represents a net condensation rate for a pure horizontal motion. In a similar way we can define

$$\mathcal{C}_z = w\left(\frac{\partial p_v}{\partial z} - \frac{p_v}{p_d}\frac{\partial p_d}{\partial z}\right), \tag{4-27}$$

where $\mathcal{C}_z$ represents now a net condensation rate for a pure vertical motion. From (4-26) and (4-27) is clear that the net condensation rate corresponds to the sum of the contributions for a pure horizontal motion and a purely vertical motion, i.e., $\mathcal{C} = \mathcal{C}_x + \mathcal{C}_z$.

By using Dalton's law $p = p_d + p_v$ and (4-27) in (4-25) after some algebra

$$u\frac{\partial p}{\partial x} = \frac{p_d}{p_v}(\mathcal{C}_z - \mathcal{C}) + u\frac{p}{p_v}\frac{\partial p_v}{\partial x}. \tag{4-28}$$

This equation shows that the horizontal pressure gradient is linked to the difference between the net condensation rate of a purely vertical motion $\mathcal{C}_z$ and the real condensation rate $\mathcal{C}$, plus a term proportional to the horizontal pressure gradient of water vapor.

To obtain the continuity equation that MGH proposes we must restring ourselves to study the continuity equation for the regions where the water vapor is saturated. We will use p_v^* to represent the saturated vapor pressure. MGH also imposes the horizontal gradients of temperature to be zero ($\partial T/\partial x = 0$) and being the saturated vapor pressure a function of temperature only, due to the Clausius-Clapeyron equation, this implies that $\partial p_v^*/\partial x = 0$. MGH justifies these assumptions by saying that in the real atmosphere this corresponds to a constant relative humidity in a weak temperature gradient approximation (See for instance the interactive discussion of Makarieva et al. (2013b). We will refer to this discussion in the following as MG13D). Under these constraints (4-28) reduces to

$$u\frac{\partial p}{\partial x} = \frac{p_d}{p_v^*}\left(\mathcal{C}_z^* - \mathcal{C}\right), \tag{4-29}$$

which is equivalent to (8) of Gorshkov et al. (2012) and where $\mathcal{C}_z^*$ represents (4-27) evaluated for the saturated case. This equation represents a limited version of (4-28) under the mentioned approximations.

It is important to remark that both (4-28) and (4-29) describes the restrictions on the pressure and velocity fields for a given condensation rate, but these equations by themselves cannot say anything about $\mathcal{C}$. Therefore, to completely resolve the fluid fields, we need to use restrictions that come from using the other primitive equations, like the energy conservation and the equations of motion together with boundary conditions.

Another way to proceed is to consider a reasonable value for $\mathcal{C}$ that might come from empirical observations. This latter approach is the one followed by MGH. Although works like Makarieva et al. (2013b) and Gorshkov et al. (2012) do not precisely state this idea, the discussions like MG13D suggest that MGH proposes that, for a parcel undergoing adiabatic condensation in a hydrostatic atmosphere, the condensation rate must be: (a) Proportional to the amount of condensing vapor, (b) proportional to the vertical velocity w and (c) proportional to the degree by which the vertical distribution of vapor deviates from equilibrium. These three points together allows MGH to propose

$$\mathcal{C}_{MGH} = w\left(\frac{\partial p_v^*}{\partial z} - \frac{p_v^*}{p}\frac{\partial p}{\partial z}\right), \tag{4-30}$$

which expresses the condensation rate as "the difference between the total change of vapor density with height and the density change caused by adiabatic expansion" (Makarieva et al., 2013b).

Using (4-30) in (4-29), yielding after some algebra

$$u\frac{\partial p}{\partial x} = w\left(\frac{\partial p_v^*}{\partial z} - \frac{p_v^*}{p}\frac{\partial p}{\partial z}\right) = \mathcal{C}_{MGH}, \tag{4-31}$$

which is the continuity equation proposed by MGH. From this equation this hypothesis is capable under reasonable assumptions to calculate the power of the Hadley cell (e.g., Makarieva et al., 2013b) and to apply this equation to study the pressure profiles in hurricanes (e.g., Makarieva and Gorshkov, 2009b; Makarieva et al., 2015). This interesting applications of MGH allows to this hypothesis to still receive some attention despite all the confrontations (A. Makarieva, personal communication, 2017).

The proposal of (4-31) received numerous criticism. For example, is important to highlight the comments of N. Stokes and I. Held in MG13D. Their main critique comes from the fact that $\mathcal{C}_{MGH}$ resembles (4-27) but replacing p_d by p. Despite the similarities, there is a clear difference between $\mathcal{C}_{MGH}$ and $\mathcal{C}_z^*$ given by

$$\mathcal{C}_{MGH} = \frac{p_d}{p}\mathcal{C}_z^*. \tag{4-32}$$

Since for moist air $p_d < p$ it is clear that $\mathcal{C}_{MGH} < \mathcal{C}_z^*$ and by comparing with (4-25) and (4-27) the proposal of MGH seems to correspond to consider a condensation rate that ignores the contribution of the horizontal motion to condensation and is less than the rate for a purely vertical motion.

Nevertheless, as we pointed out in the previous section, (4-29) only offers restrictions on the pressure and velocity fields for a given condensation rate, but it does not provide its value. Therefore, from a mathematical point of view (4-31) corresponds to a valid assumption based on the supposition that $\mathcal{C} = \mathcal{C}_{MGH}$. This point was used by MGH to defend its proposal because, as MG13D states, (4-31) does not violate the continuity equation. However, a valid mathematical assumption does not mean that this supposition is physically valid and therefore the necessary question to ask is: Is $\mathcal{C}_{MGH}$ a reasonable guess for the condensation rate?

To answer this question, we need to use the restrictions to the parcels motion that come from using the other primitive equations. Therefore, we will calculate how is the condensation rate for a moist saturated parcel undergoing adiabatic condensation in a hydrostatic atmosphere.

From (4-4) with $d\eta = 0$ for adiabatic condensation processes and by using (4-23) we can obtain a restriction for the condensation rate given by the entropy equation. For simplicity, we will consider a pure vertical motion in the steady state, yielding after some algebra

$$\mathcal{C} = \frac{p_d}{\varepsilon}\frac{dr_v^*}{dt} = -\frac{\beta(R_d T)^2}{\Lambda\varepsilon^2 L_L p_v^*} w\left(\frac{\partial p_v^*}{\partial z} - \frac{\varepsilon L_L p_v^*}{\beta R_d T^2}\frac{\partial p}{\partial z}\right), \tag{4-33}$$

where $\beta = \rho_d(c_{pd} + r_v^* c_{pv})$ and we used the Clausius-Clapeyron equation given by (Emanuel, 1994, Eq. (4.4.11))

$$dp_v^* = \frac{\varepsilon L_L p_v^*}{R_d T^2} dT. \tag{4-34}$$

From (4-30) is clear that MGH is neglecting the effects of horizontal convergence on condensation, but this omission is justified because vertical motions are the primary cause for condensation in the Earth's atmosphere. Nevertheless a simple comparison of (4-30) with (4-33) allows us to realize that the condensation rate that MGH proposes does not correspond to the condensation rate for a parcel condensing adibatically its water vapor due to vertical motion.

We can calculate the difference between the condensation rate given by MGH and the adiabatic one given by (4-33). We will assume for this comparison a hydrostatic pressure gradient pressure gradients, as MGH does. This corresponds to consider that vertical accelerations are negligible which is a good approximation for large-scale motion (Holton and Hakim, 2013, pp. 44-45), therefore the vertical gradient for p is given by

$$\frac{\partial p}{\partial z} = \rho g, \tag{4-35}$$

where ρ is the density of the moist air, and g is the gravity acceleration. Due to condensation, the saturated vapor pressure must follow the Clausius-Clapeyron equation given by (4-34).

Using this hydrostatic approximation, in Figure **4-1** we show an example of the comparison of the values for the condensation rate divided by the vertical velocity as given by MGH ($\mathcal{C}_{MGH}$) and the given for a saturated parcel undergoing adiabatic vertical motion ($\mathcal{C}_{as}$). The figure plots the case where $p = 1000$ hPa, and we varied the temperature in the range $-15°C \leq T \leq 30°C$. From this figure is clear that the magnitude of the condensation rate given by MGH is higher than the one by the adiabatic condensation process. The rates are approximately equal only at the low temperatures, where processes that we neglected in our equations become important like the production of ice. For the range of temperatures from 17°C to 30°C the differences between both rates are in the range between 5% to 16%. For lower pressures stronger differences can be found, for example, for $p = 500$ hPa the differences in the range of 7°C to 30°C are in the range between 5% to 60% (not shown).

Figure **4-1** shows a very interesting behavior: although the MGH's assumption about the condensation rate is wrong, it predicts a condensation rate close to the one for an adiabatic process for surface pressures in a reasonable range of temperatures. This behavior explains why the estimations of the power of the atmospheric circulation and the profiles for tropical cyclones according to MGH, seems to agree with the observations but it is clear that the MGH results look right but not for the right reasons.

Also note, that for the estimation of the condensation rate we used all the primitive equations. We used the continuity equation and the state equation to define the restriction of $\mathcal{C}$ on the fluid fields; we used the entropy and the Clausius-Clapeyron equations to define how $\mathcal{C}$ is in terms of the pressure gradients; and we used the equation of motion represented by the hydrostatic approximation to estimate the value of $\mathcal{C}/w$. This agrees with the well-known fact that to describe the fluid fields, it is necessary to include all the dynamical equation including the momentum, energy conservation, and the Clausius-Clapeyron equation to have a complete set of equations to solve the air motion.

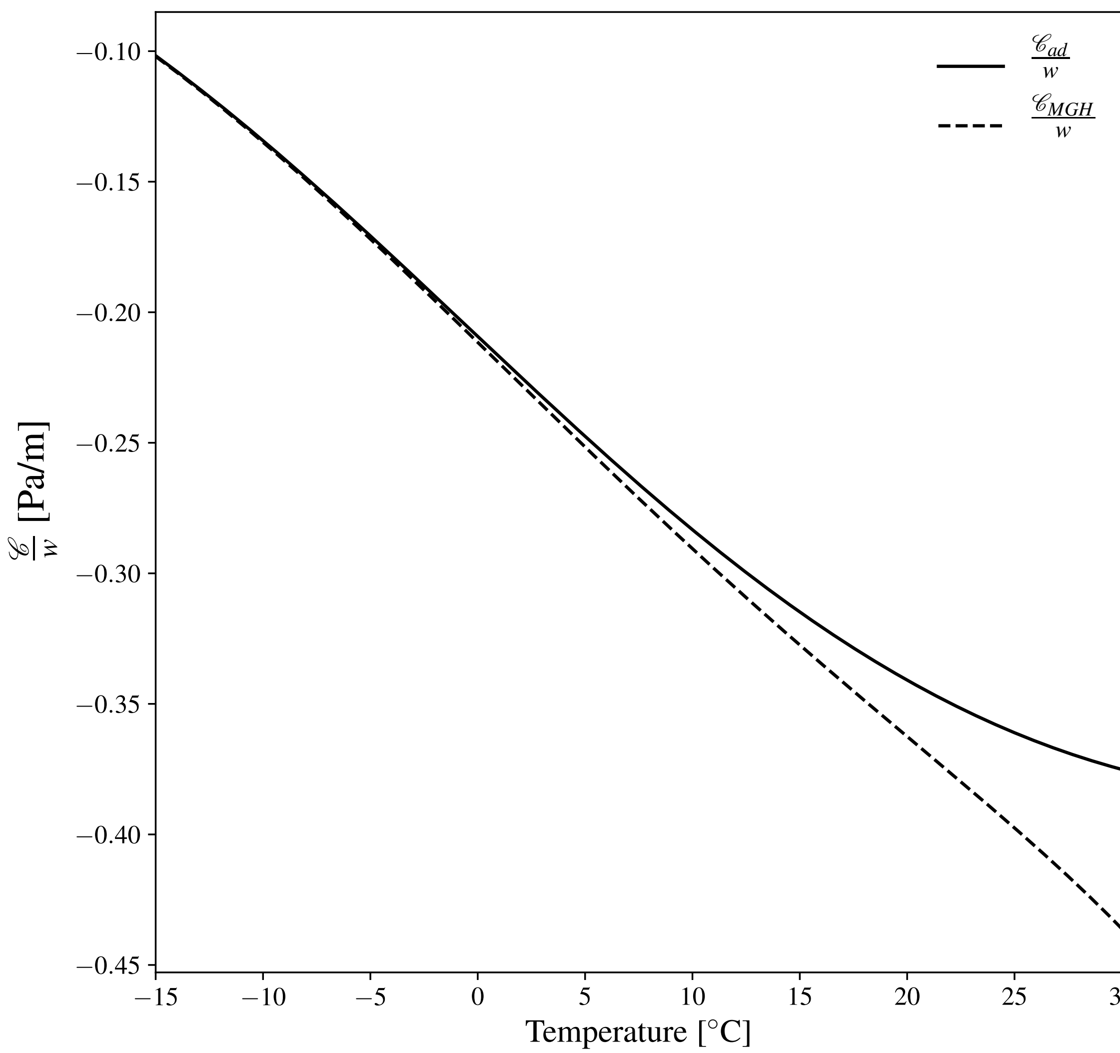

Figure **4-1**.: Condensation rate divided by the vertical velocity for a saturated parcel undergoing adiabatic vertical motion ($\mathcal{C}_{ad}/w$) and for MGH ($\mathcal{C}_{MGH}/w$). The total pressure is set to $p = 1000$ hPa and the temperature varies in the range $-15°C \leq T \leq 30°C$.

Chapter 5.

Conclusions

How forests attract rain, i.e., the question of the hydrological and meteorological role of forests is an important puzzle in the current understanding of the Earth's system complex interactions. Observations suggest a connection, but they do not explain what are the causes or the mechanisms. Numerical models, on the other hand, are evolving to include as much as our current understanding of these process allows, but they are seriously limited to resolve numerically the scales needed for a complete representation of the physics of these processes. Therefore numerical models use parameterizations to represent these processes making difficult or almost impossible the job to identify the mechanisms that explain the interactions between biota and climate.

The idea of finding a simple but a powerful physical mechanism to describe the regulation of climate by forest is undoubtedly a fascinating research goal. This goal motivated the proposal of the M-G hypothesis that we described in Chapter 2. This hypothesis proposes that a previously unstudied "evaporative force" f_e, associated with the condensation of water vapor in the gravitational field of the Earth, drives the atmospheric circulation, transporting moist air from areas of weak evaporation to regions where evaporation is intense. The high capacity of forests to sustain high evaporation rates allows them to use this force to suck in moist air from moisture sources like the oceans, providing a physical mechanism that explains how forests regulate their precipitation.

The basis of the hypothesis seems to be physically coherent, attracting the attention of the academic community, in particular on those interested in a better understanding of the interaction among the forests, the hydrological cycle, and the atmospheric circulation. On the other hand, this proposal is in an apparent confrontation with the standard theory, where the buoyancy force is the dominant driver of atmospheric motions. The standard theory uses well-established physics principles, and it is a very active area of research with thousands of experts working on questions related to different aspects of the atmosphere's interactions. The question of how is possible that in such an active field there is a space for an ignored fundamental property created serious doubts about the correctness of this hypothesis. Nevertheless, a solid physical explanation of where are the flaws in the M-G hypothesis argumentations was needed. The main result of this work is to provide such explanation.

In Chapter 3, we studied the standard view of the process of adiabatic/pseudo-adiabatic condensation, and we calculated the contribution of condensation to the buoyancy force of a moist saturated parcel ascending and condensing in the atmosphere. This chapter presents to the reader the necessary knowledge to start studying the role of condensation in the Earths atmosphere and introduce the reader coming from other disciplines to the notation and arguments used in the literature.

In Chapter 4 we present a careful analysis using the specific entropy equation for a parcel of saturated moist air to write an equation for the net buoyancy force, i.e., the difference between the vertical pressure gradient force and the weight of a rising saturated parcel. We decomposed this force into contributions from the dry air and the water vapor, and showed that the contribution of the water vapor to the net buoyancy force is equal to f_e, the proposed force by M-G. However, this force is exactly canceled by a component of the dry part. Therefore we demonstrated that not only is f_e not a new force, but it also plays no role in atmospheric dynamics.

Although this demonstration is sufficient to show that MGH is incorrect, additional analysis reveals further serious problems:

1) M-G presents two different definitions for the evaporative force, of those the one with visible physical errors is the most used in the M-G related literature.

2) M-G violates Newton's third law. If condensation creates a force f_e on the water vapor, this force is transferred to the dry air via molecular collisions. Through Newton's third law, there must appear an equal and opposite force exerted by the dry air on the moisture. Therefore, the net force on the air mixture must be zero, and it is not possible to explain atmospheric dynamics as proposed by M-G using f_e, due to this violation of Newton's laws.

3) Another confusion of M-G is evident from their assertion that the ascending and descending components of work done by the buoyancy force cancel, resulting in a significant decrease in the energy released by atmospheric circulations. We showed using the Kelvin circulation theorem that the latent and sensible heating injected into the circulation at low levels by surface fluxes have a net positive effect on the circulation. The importance of evaporation to the maintenance of atmospheric circulations is thus clear but it does not have to coincide with locations of strong precipitation. The explanation for the distribution of precipitation over regions such as the Amazon must be sought elsewhere, perhaps by studying the full water budget for such regions.

4) The continuity equation of M-G is based on an assumption for the condensation rate in the Earth's atmosphere. We showed that the rate proposed by M-G does not corresponds to the condensation rate for an adiabatic process, contrary to what M-G assumes. Although the rated proposed by M-G is close to the adiabatic one for surface pressures and normal temperatures on the surface, it does not correspond to a correct assumption for this rate, and therefore the relationship between the horizontal pressure gradients and the power of their new evaporative force is a disproportionate claim. Mass continuity alone cannot control the dynamics. It is well-known that to define a fluid dynamical system it is necessary to include all the dynamical equations including the momentum and energy conservation, and the restrictions of mass and energy conservation and the Clausius-Clapeyron equation to have a complete set of equations to solve the air motion.

From all these analyses, the main conclusion of this work is that the physical basis of the M-G hypothesis is wrong. The evaporative force thus has no fluid dynamical consequences, and the whole superstructure of M-G collapses due to a failure at its foundation. This collapse is not due merely to a failure to conform to classical ideas about geophysical fluid dynamics, but to the violation of a bedrock physical principle, Newton's third law. We also showed that the arguments of this hypothesis are plagued with numerous misconceptions and severe physical mistakes that obscured the discussion.

We agree with the M-G hypothesis that forests play a significant role in the global circulation, but this role is still poorly understood. We think, coinciding with the conclusions of Meesters et al. (2009), that the proposal of theoretical mechanisms to simplify the physics of these complex interactions between forests and climate is an interesting and valuable approach to advance the understanding of the climate system, but the proposed mechanism must be based on sound physics principles and by substantial observational evidence.

Final Thoughts

> *"During the course of proof, we must be just as diligent in seeking data contrary to our hypothesis as we are in ferreting out data that may support it. Let us avoid excessive attachment to our own ideas, which we need to treat as prosecutor, not defense attorney. Even though a tumor is ours, it must be removed."*
> **Ramón y Cajal (2004)**

The well-known mathematician G. Pólya proposed that to solve a problem we need to follow four steps:

> *"First, we have to understand the problem; we have to see clearly what is required. Second, we have to see how the various items are connected, how the unknown is linked to the data, in order to obtain the idea of the solution, to make a plan. Third, we carry out our plan. Fourth, we look back at the completed solution, we review and discuss it"* Pólya (2014).

This chapter aims to address this fourth step. The previous chapters described a problem and how we solved it. Now we will look back to the solution and the path that led to it to see what we can learn from this work.

This thesis started with a question, the question that gave the name of this work: Is the hypothesis of the condensation-induced atmospheric dynamics a new theory of the origin of the winds? which is paraphrasing Makarieva et al. (2013b). They also started with a question: "Where do winds come from?" where they explained in detail the physical principles of what they called the hypothesis of the condensation-induced atmospheric dynamics or also known as the Biotic Pump hypothesis. As this thesis describes, this hypothesis created great interest and expectations in a group of scientist that support these ideas with high hopes in what this hypothesis has to offer, especially to understand the interaction between the hydrological cycle and the atmospheric circulation. Another group of experts vigorously criticized this proposal and expressed serious doubts about the physical correctness of these ideas.

In our first approximation to the hypothesis, the concept of a physical mechanism where condensation over forest suck in air to maintain the precipitation inland seemed convincing and the arguments appeared to be physically correct. We have to admit that at first, we were eager to find an affirmative answer to our question. We started exploring if this hypothesis could be used to explain a complex phenomenon like the MJO and to create models that included the force proposed by the hypothesis, but very soon we realized that resolving the doubts about the physical argumentation was a necessary and more important step before proceeding further with the hypothesis and its applications.

As G. Pólya also pointed:

> *"Trying to find the solution, we may repeatedly change our point of view, our way of looking at the problem. We have to shift our position again and again. Our conception of the problem is likely to be rather incomplete when we start to work; our outlook is different when we have made some progress; it is again different when we have almost obtained the solution"* Pólya (2014).

We changed our point of view, and we decided to move forward with a profound and skeptical analysis of the physics of the hypothesis.

The physical arguments of the hypothesis are highly theoretical and, as we showed in this work, the main problem to understand the flaws in the arguments are the subtle but significant errors in the arguments of the hypothesis. This study implied a good understanding of the physics and thermodynamics of the atmosphere, that although are well-known subjects, sometimes they are susceptible of misinterpretations as J. H. Dutton points:

> *"[Thermodynamics] is often a confusing subject, mainly because written explanations have generally preceded the equations, and such verbal descriptions of thermodynamics experience are subject to misinterpretations"* (Dutton, 2002)

Our conclusions, although they may seem obvious to some, point to a deeper problem in the scientific debates of this kind of theoretical topics. There is a need to revise the foundations of the standard theory and make ourselves questions to look for a better understanding of the theory and its value to study the different phenomena that we intend to comprehend. On these grounds, we can see a positive aspect of the study of M-G hypothesis: it opened the debate about important topics that were relegated to textbooks only. The difficulty that the different participants in the discussion showed in explaining where the flaws in the arguments were, clearly points to the need for a better understanding of the physics foundations of the theory of atmospheric dynamics. For this reason, our approach, although it might seem simplified, presented an analysis from the most simple point of view: transform the question into a simpler problem where we can play with the physical principles to understand the fundamental concepts of it. This approach is in our opinion the best methodology to expose the flaws in the arguments presented in scientific debates where there are different and contradicting theoretical points of view.

To finish this work, we can do an interesting thought experiment. We will ignore for a moment that we know that the M-G hypothesis is based on wrong physics and let us ask ourselves: Would we have come to the same conclusion if we had followed another path?

Some of the most promising or tempting ways to test the M-G hypothesis would be experimentation, observation, and modeling. Let us freely speculate something about each of this options

- **Experimentation:** A possible way to test the existence and importance of the evaporative force is by constructing and experiment that under controlled conditions reproduced the circumstances for the appearance of the evaporative force in the Earth's atmosphere. This approach would be by far the best test of the scientific hypothesis but, as we know, condensation in the atmosphere occurs in scales bigger that what we can reproduce in a laboratory; The vertical scale is of the order of the height of the troposphere and the horizontal scale might be several orders of magnitude that distance. Therefore it seems difficult to propose and effectuate an experiment where we can control the conditions on such a big scale, at least with our present technology.

- **Observations:** Studying observations of how precipitation varies in forested zones seem at first an interesting approach to study the M-G hypothesis. In fact, this observations approach was the main motivation presented by their authors to propose their ideas. Nevertheless, as we already explained in this work, observations suggest connections between different phenomena and systems, but they do not offer an explanation of what are the causes or the mechanism.

- **Modeling:** Constructing numerical models that include this evaporative force seems one of the most promising of the three options. Nevertheless, this approach presents some difficulties that we must to account. First, and most important, to construct numerical models we need to have a complete formulation of the dynamical equations where the evaporative force is present. The papers that propose the evaporative force present the problem in a stationary motion, and therefore they do not provide the dynamics needed for studying the time evolution of the atmospheric system. Let us suppose for a moment that we have such a dynamical formulation; the next problem would be that a numerical model needs to parameterize the scales that it can not solve numerically. These parametrizations introduce considerable uncertainty about the conclusions we can get from our model; are our results the direct consequence of the introduction of the evaporative force? or are they the result of the parametrizations we introduced? It would be hard or impossible to know even by comparing with the results of numerical models that do not include the evaporative force.

From these ideas, it seems that these approaches would not provide a definitive answer to our initial question and therefore doubts about this hypothesis would continue unanswered. Our approach showed that the hypothesis did not survive the theoretical test and efforts to understand the forest regulation of climate must be focused on physically sound proposals.

As our work showed, the circulations described by the M-G hypothesis can be understood by the standard theory if we concentrate on the importance of condensation. Condensation is, therefore, an important process in the Earth's atmosphere and, as the M-G hypothesis suggest, it plays an important function in the link between forests and the atmosphere. Future research efforts should focus on understanding how forests and their biology can affect and regulate condensation, improving our understanding of the relationship between the hydrologic cycle and the atmospheric circulation. Condensation might be the key process to bring closer disciplines like hydrology, meteorology, and biology in a consistent picture that will improve our understanding of the climate system and the physics of the future atmospheric numerical models. One of such possible approaches can be the one proposed by Wright et al. (2017) with their "Deep convection moisture pump."

Finally, we can answer our initial question: **Is the hypothesis of the condensation-induced atmospheric dynamics a new theory of the origin of the winds?**

No, but this negative answer is an invitation to persevere in the fascinating problem of the forest regulation of climate but now from sound theoretical foundations.

Appendix A.

On the Relative Density of Clouds[1]

A.1. Introduction

Pelkowski and Frisius (2011), hereafter PF11, pointed out the importance of the proper definition of densities in atmospheric thermodynamics. They showed that precise answers on how densities of clouds compare with those of moist or dry air are difficult to find. PF11 explored the literature in search of answers to this question. They found that authors generally agree that, under the same pressure and at the same temperature, moist air is less dense than dry air, but when dealing with cloud air some disagreements between authors were found. In particular, PF11 contrast the conclusions by Dufour and Van Mieghem (1975) who assert that under the same circumstances, cloudy air is less dense than moist air and by Bohren and Albrecht (1998) that state the opposite conclusion. PF11 found that cloudy air is denser than moist air and that in general cloudy air is less dense than dry air, only being denser when the mixing ratio of the condensate content is higher than about 60 percent of the water vapour mixing ratio in the cloudy mixture. The work of PF11 represents a very rigorous way of defining densities in atmospheric thermodynamics, but it is limited only to comparisons of parcels at the same pressure, temperature, and water vapor mixing ratio.

It is worthwhile to have a closer look at the complex question of densities of atmospheric air masses from a simpler but general perspective. This perspective can be presented from the definition of virtual temperature, defined as the temperature that dry air would have if its pressure and density were equal to a given sample of moist air (Glickman, 2000), and the density temperature, a generalization of this concept by Saunders (1957) to include condensed water, sometimes also called virtual temperature with respect to cloud air. Care is needed because some derivations of density temperature presented in the literature do not give proper physical justifications. In

[1]This appendix corresponds to Jaramillo and Mesa (2017a)

particular, there are two different possible definitions of density temperature depending on the size of the hydrometeors. For sizes of the order $\mathcal{O}(< 1\ \mu\text{m})$, due to the Brownian forces exerted by moist air, hydrometeors are in suspension. This formulation of density temperature is common in meteorological literature, but the different authors consider the effect of hydrometeors on density, neglecting their effects on pressure with no justification. We will show that a correct way to define such effects comes from considerations of colloidal physics and the definition of osmotic pressure, but for the normal conditions of Earth's atmosphere these osmotic pressures in clouds can be ignored. When the hydrometeor size is higher than 1 μm, hydrodynamic forces prevail over Brownian forces and the hydrometeors cannot be considered in suspension. Under those conditions, a dynamical definition of density temperature is needed such as the one proposed by Monteiro and Torlaschi (2007).

Using standard notation from textbooks such as Emanuel (1994) we present a simpler but general expression to compare cloud densities and we test this approach comparing our conclusions with those of PF11. In addition, we consider some other cases not included by them.

A.2. Mixtures of moist and cloudy air

Following Emanuel (1994), we will consider cloudy air as a mixture of dry air, water vapour and hydrometeors (liquid or solid) corresponding to the water condensate content. For a non-precipitating parcel, air is filled with cloud droplets that may be considered to be falling at their terminal velocities, to a good approximation. From this point of view, clouds may be considered to be in suspension (Emanuel, 1994). As long as these assumptions are valid, hydrometeors behave as colloidal particles dispersed in a fluid of moist air undergoing perpetual irregular motions due to chaotic collisions with the molecules of the medium. To a good approximation, those motions can be considered as Brownian motions, an idea that has been used extensively to study colloidal, cloud and aerosol physics (e.g., Russel et al., 1989; Straka, 2009; Pruppacher and Klett, 1996). Since the work of Einstein (1905) on Brownian motion it is well known that colloidal particles dispersed in a fluid of smaller molecules behave as a thermodynamic system (Frey and Kroy, 2005; Brady, 1993). These colloidal particles exert a finite, albeit small, osmotic pressure on the solution. An entirely mechanical definition of this osmotic pressure is also possible by studying the hydrometeors' effects from an hydrodynamic approach as shown by Brady (1993).

Works like that of Saunders (1957) or classic textbooks such as those by Stull (1988), Emanuel (1994) and even PF11 came to the conclusion that hydrometeors have effects on density but disregard their effects on pressure without proper justification. We will show that the osmotic pressure exerted by the hydrometeors is negligible and the procedures commonly presented in the above references can be justified if one considers hydrometeors as a pseudo-gas. This was pointed out by Monteiro and Torlaschi (2007). The concept of osmotic pressure is common in physics (e.g. Landau et al., 1967) and complex fluids literature (e.g. Brady, 1993). Nevertheless, as we will show, the consideration of a pseudo gas does not change any of the current well-known results, although it adds a more consistent physical conceptual foundation.

The total pressure of the cloud parcel is the sum of the partial pressures of dry air p_d and water vapour p_v that can both be treated as ideal gases. In addition to these pressures we have to add the "partial" osmotic pressure of the hydrometeors p_{osm}. This "partial" osmotic pressure is given by Van't Hoff's formula $p_{osm} = nkT$ (Landau et al., 1967; Brady, 1993), where n is the number density of hydrometeors, $k = 1.38064852 \times 10^{-23}\,\mathrm{JK^{-1} molecule^{-1}}$ is the well-known Boltzmann's constant and T is the absolute temperature. Note that this form of the osmotic pressure is similar to the pressure of an ideal gas composed by "droplet molecules". Therefore, the total pressure can be written as:

$$p = p_d + p_v + p_{osm} = \rho_d R_d T + \rho_v R_v T + \rho_c R_c T, \tag{A-1}$$

where $R_{d,v} = R/M_{d,v}$ are the gas constants for the dry air and water vapour, $R = kN_a$ the universal gas constant with units $\mathrm{JK^{-1} mol^{-1}}$, $N_a = 6.02 \times 10^{23}\,\mathrm{molecule\,mol^{-1}}$ is the Avogadro number, $M_v = 0.01806\,\mathrm{kg\,mol^{-1}}$ is the molar mass of water vapour and $M_d = 0.02897\,\mathrm{kg\,mol^{-1}}$ is the apparent molar mass of dry air (Wallace and Hobbs, 2006). For the "partial" osmotic pressure by hydrometeors we define $\rho_c = nM_c/N_a$ as the partial density of the hydrometeors and $R_c = R/M_c$ where M_c is the mass of a mole of hydrometeors, similar to the definition of the molar mass for ideal gases and will depend on the droplet size. Note that Eq. (A-1) resembles Dalton's law of partial pressures of three ideal gases: dry air, water vapour and a hydrometeor pseudo-gas.

If we define the total density as $\rho = \rho_d + \rho_v + \rho_c$, we can rewrite Eq. (A-1) as

$$p = \rho R_d T \delta_r, \text{ with } \delta_r = \frac{1 + r_v/\varepsilon_v + r_c/\varepsilon_c}{1 + r_v + r_c}, \tag{A-2}$$

$r_{v,c} = \rho_{v,c}/\rho_d$ are respectively the mixing ratios of water vapour and water condensate, and $\varepsilon_{v,c} = R_d/R_{v,c}$ the ratios of the dry air gas constant to the gas constant of water vapour or condensate. In this equation the terms $r_{v,c}/\varepsilon_{v,c}$ in the numerator can be interpreted as a measure of the contribution of the pressure of water vapour or the "osmotic" pressure to the total pressure of the gas, where the dominating contribution comes from the dry air pressure. For Earth's atmosphere r_v is observed to be generally less than 0.04 and r_c is typically of the order of $\mathcal{O}(10^{-3})$ (Emanuel, 1994). Taking $R_d = 287.0\,\mathrm{JK^{-1}kg^{-1}}$ and $R_v = 461.51\,\mathrm{JK^{-1}kg^{-1}}$, this gives $\varepsilon_v = R_d/R_v = 0.622$ and $r_v/\varepsilon_v \sim \mathcal{O}(10^{-2})$. To calculate $\varepsilon_c = R_d/R_c$ we must consider that for hydrometeors M_c will depend on the droplet size. We can estimate an order of magnitude of M_c using a typical droplet size for Earth's clouds. Observations of the size distribution of cloud droplets near stratus cloud base show high number concentrations of small droplets

< 2.5 μm radius (Akagawa and Okada, 1993). For a cloud with droplets with this radius we estimate $M_c = 4\times 10^{10}\,\mathrm{kgmol^{-1}}$, $R_c \sim 2\times 10^{-10}\,\mathrm{JK^{-1}kg^{-1}}$ and $\varepsilon_c \sim \mathcal{O}(10^{12})$, then $r_c/\varepsilon_c \sim \mathcal{O}(10^{-15})$ compared with $r_v/\varepsilon_v \sim \mathcal{O}(10^{-2})$. From these orders of magnitude, it is clear that for typical atmospheric conditions $r_v/\varepsilon_v \gg r_c/\varepsilon_c$ and the term r_c/ε_c can be neglected, allowing us to say that $p_{osm} \approx 0$ and explaining why the osmotic pressure has not appeared in the meteorological literature. Another way to argue the smallness of the contribution of the condensate to the total pressure is directly from the standard kinetic expression for the pressure, $p = NRT$, with N the molar density. The partial pressure of the condensate is very small because N for the condensate is very small in comparison with the air molar density. The two arguments are equivalent; the first one is due to the common presentation of the ideal gas law using a specific gas constant and the second one is clearer because the ideal gas law does not depend on the molar mass. Then, δ_r can be approximated as

$$\delta_r \simeq \frac{1 + r_v/\varepsilon_v}{1 + r_v + r_c} \approx 1 + r_v\frac{1-\varepsilon_v}{\varepsilon_v} - r_c. \tag{A-3}$$

The factor δ_r can be interpreted as a correction factor to the pressure of a parcel of dry air with the same total mass of the mixture that is usually applied to either the gas constant or to the temperature. We can define $R_a = R_d\delta_r$ as an effective gas constant (Emanuel, 1994, Eq. (4.2.9)) or $T_\rho = T\delta_r$ as a density temperature (Saunders, 1957; Deardorff, 1980; Stull, 1988; Betts and Bartlo, 1991; Emanuel, 1994) and we can write the equation for the total pressure either as

$$p = \rho R_a T \quad \text{or} \quad p = \rho R_d T_\rho. \tag{A-4}$$

If $r_c = 0$ the density temperature T_ρ reduces to the well-known definition of virtual temperature (Emanuel, 1994, Eq. (4.3.1))

$$T_v = T\frac{1 + r_v/\varepsilon_v}{1 + r_v}. \tag{A-5}$$

This definition applies as long as the cloud droplets may be considered to be in suspension. For cloud droplet sizes greater than 1 μm the individual bombardment by microscopic gas molecules will have little effect and Brownian forces become insignificant (Seinfeld and Pandis, 1998; Brady, 1993) and a new definition of density temperature should come from hydrodynamic considerations. Monteiro and Torlaschi (2007) redefined density temperature from a dynamical point of view, showing that a more adequate definition of virtual temperature is the temperature that a parcel of dry air should have in order to experience the same acceleration as a parcel of cloud air and not in terms of its densities as is usually done in the literature. They started from the dynamical equations for moist air from Bannon (2002) and, performing a scale analysis, they showed that the contributions due to the acceleration and phase transitions of the condensate particles are negligible with respect to the contributions of gravity. Even though the dynamic definition is more appropriate for big droplet sizes, the approximations reduce to the same expression for density temperature that we have presented here. Therefore, the present formulation of density temperature corresponds to a general expression that takes into account densities in clouds corresponding to all ranges of droplet sizes, as long as interactions between hydrometeors can be ignored.

A.2.1. A general expression to compare air densities

We can use Eq. (A-2) to compare two air parcels, A and B. PF11 restrict the comparison to the case of equal pressure, temperature and water vapour mixing ratio. But in general, in atmospheric thermodynamics, parcels are usually compared at equal pressure only, therefore the most general expression to compare the densities of parcels at equal pressure is

$$\rho_A = \rho_B \frac{T_B(1 + r_{v,B}/\varepsilon_v)(1 + r_{v,A} + r_{c,A})}{T_A(1 + r_{v,A}/\varepsilon_v)(1 + r_{v,B} + r_{c,B})} = \rho_B \frac{T_{\rho,B}}{T_{\rho,A}}, \tag{A-6}$$

where we have added an index to each term, to distinguish the properties of parcel A or B. From this equation, it is evident that the most general comparison of two air parcels comes from the ratio between their density temperatures. In the next sections we will use this equation to study the cases presented by PF11 and some others not considered by them.

A.2.2. Density of cloudy air compared with density of dry air

The interpretation of the correction factor δ_r in the definition of either an effective gas constant or of a density temperature can be misleading when facing the question of whether cloudy air is denser than dry air. First, we compare a cloudy air parcel with a parcel of dry air. For this comparison parcels of cloudy and dry air are taken at the same temperature and pressure. From Eq. (A-6), the density of cloud air compared with the parcel of dry air is

$$\rho_{cloud} = \rho_d \frac{1 + r_v + r_c}{1 + r_v/\varepsilon_v}. \tag{A-7}$$

From this equation, it is clear that the density of a cloud parcel will be less or greater than that of a dry air parcel depending on whether the fraction is greater or less than 1. In particular, if we compare moist air ($r_c = 0$) with dry air ($\rho_{cloud} = \rho_{moist}$ in this case), the value of the denominator is greater than the value of the numerator, giving the well-known result that moist air is always lighter than dry air under the same pressure and temperature. Returning to the general case, the condition that defines whether cloudy air is lighter or denser than dry air can be found by studying r_c^*, the value of r_c for which both densities are equal

$$r_c^* = r_v \frac{1 - \varepsilon_v}{\varepsilon_v} = 0.608 r_v. \tag{A-8}$$

Eq. (A-8) determines r_c^* in terms of r_v for the neutral case, that is when the density of cloudy air is the same as the density of dry air. If $r_c < 0.608 r_v$ the density of cloud air is less than that of dry air, and for $r_c > 0.608 r_v$ cloud air is denser than dry air. This result agrees with PF11 (see text after their Eq. (15)). They stated that in general, it is likely that clouds will be lighter than dry air since $r_c > 0.608 r_v$ is harder to find in normal cloud configurations. PF11 pointed out that this condition is conceivable in situations of cold air holding a condensed phase, like in cumulonimbus clouds with high liquid water content and low water vapour mixing ratios. This can be an important mechanism in the formation of mammatus clouds, where this form of downward convection is sometimes observed on the underside of middle- or high-based stratiform clouds, most frequently those associated with outflow from strong thunderstorms like the cumulonimbus anvils (Emanuel, 1981). The zones underneath the anvil where mammatus forms are often characterised by a strong vertical temperature gradient and strong gradient of moisture (Schultz et al., 2006). This difference in density between the cloud air with low water vapour mixing ratios and the dry air underneath might cause the cloudy parcel to be detrained into this dry sublayer, triggering the mechanism of cloud-base detrainment instability (CDI) proposed by Emanuel (1981) (see his Fig 3.). Under those conditions, condensed water in the cloud air is introduced into the dry air by mixing. Numerical simulation of mammatus showed that the CDI criterion appears to be a necessary, but not sufficient, condition for the formation of mammatus (Kanak et al., 2008). Published research on mammatus is limited and the observations do not include all the thermodynamic, moisture and dynamic measurements for a rigorous evaluation of the possible mechanism. As a result, relatively little is known about the physical characteristics that describe the mammatus and more remains to be learned about the physical processes that affect them (Schultz et al., 2006; Kanak et al., 2008).

Although PF11 focused on clouds composed only of liquid water (or ice) droplets, we want to point out that the very same principles apply to the study of densities of clouds composed of volcanic ash or other particulate matter as in air pollution, where the particles composing the ash or pollution play the same role as the gas of water drops. For these kinds of clouds r_c takes the role of ash or pollution mixing ratio and the condition $r_c > 0.608 r_v$ is easily fulfilled by these types of clouds, especially for dry air conditions. This may constitute a serious risk to air quality in big cities where the probability of high concentration of pollution is common on dry days and as a result, the polluted cloudy air concentrates in the lowers parts of the atmosphere (e.g. Dickerson et al., 1997; Green, 1995; Smoyer-Tomic et al., 2003).

A.2.3. Density of cloudy air compared with density of moist air

Now we compare the densities of cloudy air and moist air at the same temperature and pressure. PF11 only compared parcels at the same pressure, temperature and water vapour mixing ratio. We consider the more general case where the air parcels may have different water vapour mixing ratios, where the PF11 case is only a particular case of our example. We will add an index **c** and **m** to represent properties of cloud air and moist air respectively.

Using the same strategy as for the dry air case, from Eq. (A-6) we get an expression for comparing the densities of ρ_{cloud} and ρ_{moist} at the same temperature and pressure,

$$\rho_{cloud} = \rho_{moist} \frac{(1 + r_{v,\mathbf{c}} + r_{c,\mathbf{c}})(1 + r_{v,\mathbf{m}}/\varepsilon_v)}{(1 + r_{v,\mathbf{m}})(1 + r_{v,\mathbf{c}}/\varepsilon_v)}. \tag{A-9}$$

Clearly, if the water vapour mixing ratios for both parcels are equal, as in PF11, the fraction in Eq. (A-9) is greater than 1. It follows that cloudy air is always denser than moist air under the same pressure, temperature and water vapour mixing ratio, agreeing with the conclusion of PF11.

In general when $r_{v,\mathbf{m}} \neq r_{v,\mathbf{c}}$, there are conditions on $r_{c,\mathbf{c}}$, $r_{v,\mathbf{c}}$ and $r_{v,\mathbf{v}}$ that determine which parcel is denser. After elementary algebra the condition for the neutral value of $r_{c,\mathbf{c}}$ for equal densities is

$$r^*_{c,\mathbf{c}} = (1 - \varepsilon_v)\frac{r_{v,\mathbf{c}} - r_{v,\mathbf{m}}}{\varepsilon_v + r_{v,\mathbf{m}}} \approx 0.608(r_{v,\mathbf{c}} - r_{v,\mathbf{m}}). \tag{A-10}$$

Note that Eq. (A-10) reduces to Eq. (A-8) for the dry air case when $r_{v,\mathbf{m}} = 0$. The last approximation in Eq. (A-10) comes from the fact that in general ε_v is bigger than $r_{v,\mathbf{m}}$. From Eq. (A-9), the fraction increases with $r_{c,\mathbf{c}}$, which implies that cloudy air is less dense than moist air if $r_{c,\mathbf{c}} < r^*_{c,\mathbf{c}}$ and denser if $r_{c,\mathbf{c}} > r^*_{c,\mathbf{c}}$. Moreover, the difference $r_{v,\mathbf{c}} - r_{v,\mathbf{m}}$ imposes a very important restriction: if $r_{v,\mathbf{c}} < r_{v,\mathbf{m}}$ then $r^*_{c,\mathbf{c}} < 0$ which means that under this configuration cloudy air can only be denser than moist air because $r_{c,\mathbf{c}} > 0$.

To consider the effect of $r_{v,\mathbf{c}}$ and $r_{v,\mathbf{m}}$ one can consider the partial derivative of $r^*_{c,\mathbf{c}}$ with respect to each of those variables. Clearly $r^*_{c,\mathbf{c}}$ increases with $r_{v,\mathbf{c}}$, which means that for greater water vapour mixing ratio in the cloud, the neutral value of the mixing ratio of cloud condensate increases, making it more feasible for a cloud to be less dense than moist air. On the other hand, a simple calculation shows that $r^*_{c,\mathbf{c}}$ decreases with $r_{v,\mathbf{m}}$. Expressed in words, if the water vapour mixing ratio in moist air increases, the neutral value of the mixing ratio of cloud condensate decreases. This makes it more likely that a cloud could be denser than moist air.

Under normal atmospheric conditions, it is expected that the water vapour mixing ratio of a cloud is very close to that of the moist air surrounding it. This forces $r^*_{c,\mathbf{c}}$ to be very small and so in general cloudy air is denser than the surrounding moist air under the same pressure and temperature, but there is still a possibility that cloudy air can be lighter. In fact, if $r_{v,\mathbf{c}} - r_{v,\mathbf{m}} > 0$ there is a range of values of $r_{c,\mathbf{c}}$, from 0 to $r^*_{c,\mathbf{c}}$, where the cloud parcel is lighter than the moist parcel.

In simple words, adding droplets to an air parcel does not change the parcel's pressure. At the same time, it is evident that such an addition does increase the parcel's mass. That is why cloudy air, at the same pressure and temperature, is usually denser than moist air without droplets. Cloudy air can be lighter if and only if while adding droplets we simultaneously remove enough molecules of the heavier gas (dry air) and replace them with an equal number of lighter molecules (water vapour), where equal numbers of molecules are necessary to keep the pressure constant.

A.3. Conclusions

Air density comparisons are an important issue not clearly presented in the literature. As PF11 showed, it is even possible to find contradictory conclusions between authoritative works. Works like PF11 contribute to provide theoretical foundation to these comparisons. Although it is rigorous, PF11 only focused on comparisons of air parcels at equal pressure, temperature, and water mixing ratio. To complement this, we present a simpler and more general approach to studying air densities by means of comparing the so-called density temperatures. Previous studies usually assume that hydrometeors have effects on density but no effects on pressure. We show that considering the hydrometeors as a colloidal component of a mixture, their contribution to the total pressure comes from the osmotic pressure that resembles the ideal gas law for a gas composed of "droplets molecules". We showed that for the clouds of Earth's atmosphere this osmotic force is several orders of magnitude smaller than the contribution of dry air and water vapour and therefore can be ignored as is usually done in the literature. This conceptual clarification, although it does not change the practical definition of density temperature, sheds light on the basic concepts of atmospheric thermodynamics. We also pointed out that the validity of this definition holds as long as the cloud droplets may be considered to be in suspension. For cloud droplet sizes greater than 1 μm the individual bombardment by microscopic gas molecules will have little effect as Brownian forces become insignificant and a new definition of density temperature should come from considerations of hydrodynamics, as is done by Monteiro and Torlaschi (2007). They redefined virtual temperature as the temperature that a parcel of dry air should have in order to experience the same acceleration as a parcel of cloud air, but their scale analysis shows that this new definition coincides with the one we presented here.

Appendix B.

A Simple Model for Moist Convection[1]

The standard theory explains the closed circulations characterized by the rising of warmer, lighter air and sinking of colder, denser air and the prevalence of cross-isobar horizontal flow toward lower pressure, releasing of potential energy and conversion to the kinetic energy of the horizontal flow. These circulations are known as thermally direct circulations because they operate in the same sense as the global kinetic energy cycle. Moreover, it is also known that moist circulations render motions more vigorous that they would be in a dry atmosphere pointing to the importance of condensation in the circulation (Wallace and Hobbs, 2006, pp. 298-300). To explain this argument in a simplistic way we can use the simple model of a moist circulation described by the circuit presented in Figure **B-1**, The circuit represents the path followed by the air parcels in a closed circulation. In the lower boundary, the parcels move horizontally (DA) absorbing heat and moisture until they become positively buoyant and rise. In AB the positively buoyant parcels ascend adiabatically/pseudo-adiabatically condensing their water vapor, this is possible if we consider that the timescale of vertical motion is short compared to that of heat transfer until the parcel reaches the upper boundary. In the upper boundary the parcels move horizontally again (BC) releasing heat until they become negatively buoyant. Then in CD the parcels sink adiabatically until they reach the lower boundary completing the cycle. To study this circulation, we can use the concept of absolute circulation that is amply used in textbooks to study buoyancy driven circulations like the sea breeze problem. The absolute circulation is a macroscopic measure of rotation, where the rate of change for the absolute circulation is given by (Holton and Hakim, 2013, p. 97):

$$\frac{dC_a}{dt} = -\oint \frac{dp}{\rho}, \tag{B-1}$$

where the closed integral can be split in the branches depicted in Figure (**B-1**).

[1]This Appendix corresponds to a textbook-like approximation to the understanding of the role of condensation in the circulation of moist air. Therefore the conclusions of this analysis are very restricted and should not be extrapolated for more complex problems.

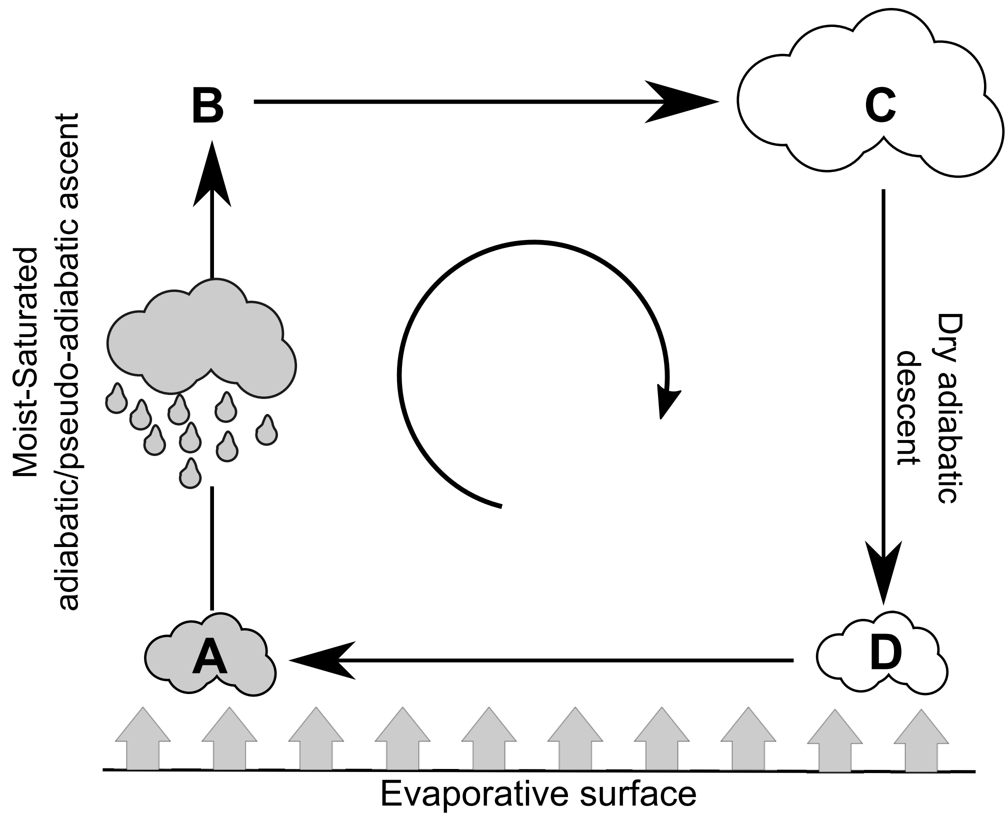

Figure **B-1**.: Circuit followed by a moist parcel during which it absorbs heat and moisture in the lower boundary (DA) and rejects heats at the upper limit (BC). The vertical motions are adiabatic/pseudo-adiabatic, where the moist parcel condensate their water vapor during the ascent (AB) and descent is dry adiabatic (CD).

The branches AB and DC of this circulation are adiabatic/pseudo-adiabatic, and Eq. (3-3) can be used to rewrite the corresponding integrals. For simplicity, we will consider that parcels follow constant pressure processes in the horizontal branches DA and BC without affecting the main conclusions at the end of this section. Therefore the integrals corresponding to the horizontal branches disappear and the equation for the rate of change of absolute circulation for the moist circulation that we described can be written as:

$$\frac{dC_a}{dt} = -c_{pd}\oint dT - \frac{L_L}{T_F R_d}\oint \frac{p}{\rho} dr_v^*. \tag{B-2}$$

For a dry circulation (no condensation) in a barotropic fluid, the density is a function of pressure only and both integrals in Eq. (B-2) are zero; The absolute circulation is conserved following the motion, which is the well-known Kelvin's circulation theorem. In general, density is not a function of pressure only and the first term of Eq. (B-2) is not zero; nevertheless, the input from the ascending branch AB will have opposite sign from the input of the descending branch CD, resulting in a significant cancellation of the net contribution of this integral to the rate of change of the absolute circulation.

For a moist circulation, the difference between adiabatic and pseudo-adiabatic ascent becomes relevant. If we have an adiabatic ascent, where all the condensate products remain inside the parcel, the contribution in the ascending branch due to condensation will cancel out the input of the descending branch due to evaporation, canceling out the net contribution of condensation to the rate of change of the absolute circulation. In simple words, for a non-precipitating system, there will be a zeroing of the contribution of condensation to the rate of change of the absolute circulation and the circulation driven by buoyancy is very constrained as the MGH affirms.

For a pseudo-adiabatic ascend instead, where all the condensate products immediately precipitate in the ascending zone, the only contribution of condensation is in the ascending branch AB, and therefore there is a not canceling contribution of condensation to the rate of change of the absolute circulation. For precipitating systems, condensation will add a significant contribution to the rate of change of the circulation, and for a barotropic atmosphere where the first integral in Eq. (B-2) is zero, the contribution of condensation will be the only one to the rate of change of the absolute circulation.

In general, we can see from these simple analyses that precipitating systems will contribute to the intensification of the moist buoyancy circulation and if in Eq. (B-2) the second integral is bigger than the first integral, condensation will drive the circulation. Also, It is evident that the air parcels will move from higher to lower pressure in all the branches of the circuit that we described. Therefore, the contribution of the horizontal branches that we neglected, given by $-\int dp/\rho ds$, correspond to an additive contribution to what we have described, justifying the simplification used. These arguments show that the cumulative potential energy related to buoyancy driven circulation is not small as the MGH suggests in especial where precipitation is present, i.e., condensation plays a significant role in atmospheric circulation.

From these basic arguments, it is clear that the standard theory explains that condensation will accelerate the atmospheric circulations in heavy precipitating systems, and for an isotropic atmosphere it will be the main contribution to low-level atmospheric circulation. The intensification of the circulation will force air to move from zones of less precipitation to zones of intense precipitation, i.e. heavy precipitating systems suck in atmospheric air, importing humidity that will help to sustain precipitation in the ascending zone maintaining the circulation. This picture explained by the standard theory is qualitatively similar to the dynamics proposed by the MGH, which attracted the attention of the academic community as a possible mechanism to understand the interaction between forests and the atmospheric circulation, without the serious physical mistakes.

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