A Study on Systems of the Global Hydrologic Cycles with Stable Water Isotopic Information

水の安定同位体比情報を用いた
地球水循環過程の解明に関する研究

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Thesis Abstract

This thesis is dedicated for a study on systems of global hydrologic cycles with the use of stable water isotopes (HDO and H$_2^{18}$O). As most of isotopic researches have provided large number of new insights into the age, origin, and pathway of water movement and other useful applications, the studies on isotopes in precipitation have been trying to understand the formation processes that make the isotopic spatio-temporal heterogeneity, and they are regarded very important not only because many isotopic studies on ground hydrology must start with them, but also because they are used as proxy for reproduction of paleo-climate.

However, because there are too many explanation variables that possibly influence the variability of precipitation isotopes, such as meteorological status, temporal and geographical conditions, etc., and because of their unclear relationships with precipitation isotopes or/and inbetween, the hydrologic processes that controls precipitation isotopes have not been quantitatively explained. This thesis therefore challenged to describe why this rain has these isotopic values, and what these isotopic values tell us.

The author first creates models that incorporate isotopic processes to quantitatively induce the mechanism of observed variability of precipitation isotopes. The model is designed to adopt additional physical processes as model components that possibly influence the variability, so that author could examine degree of the influences to precipitation isotopes. Concurrently, the author suggests a new use of water isotopes, namely “evaluation of the global hydrologic cycles by precipitation isotopes”. The framework of the evaluation was developed in this thesis. The knowledge from the inductive modeling study above improves the performance of the evaluation, and provides characteristic and interpretation of evaluated results. Furthermore, analytic investigations of the dynamic motion of atmospheric water advection are derivatively performed for finding of more direct interpretation of the isotopology (the sixth). From the second chapter through the sixth chapter, implementations and investigations of topic(s) or part of topic(s) are individually or successively described below.

The second chapter seeks to explain the causes of short-term (1–10 days) variability in precipitation isotopes, on which two types of models, i.e., Rayleigh-type models and isotope-AGCMs (Atmospheric General Circulation Models), in the past have not been able to examine clearly. A new water isotope circulation model on a global scale that includes a Rayleigh equation and the use of external meteorological forcings is developed. Transport and mixing processes of water masses and isotopes that have been neglected in earlier Rayleigh models are included in the new model. A simulation of $^{18}$O for 1998 is forced with data from the Global Energy and Water Cycle Experiment (GEWEX) Asian Monsoon Experiments (GAME) reanalysis. The results are validated by GNIP (Global network of isotopes in precipitation) monthly observations with correlation R=0.76 and a significance level >99%, and daily observations at three sites in Thailand with similar correlation and signifi-
cance. A quantitative analysis of the results shows that among three factors that cause isotopic variability, the contribution of moisture flux is the largest, accounting for 37% at Chiangmai, and 46% globally. This highlights the importance of transport and mixing of airmasses with different isotopic concentrations. A sensitivity analysis of the temporal and spatial resolution required for each variable is also made, and the model is applied to two additional datasets. The more accurate GPCP (Global Precipitation Climatology Project) precipitation dataset yields improved model results at all three observation sites in Thailand. The National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis allows the simulation to cover two years, reproducing reasonable interannual isotopic variability.

Successively, in the third chapter, a global Rayleigh-type H₂¹⁸O circulation model developed in the second chapter was forced by two reanalysis products, NCEP/NCAR reanalysis (NRA1) and the European Centre for Medium-Range Weather Forecasts (ECMWF) 15-year reanalysis (ERA15) datasets, and the GNIP database was used to validate for 1979-1993 on monthly to inter-annual scales. We find out that the model monthly reproducibility is better with ERA15, particularly in Europe; the number of sites with a statistically >99% significant correlation is 87 out of 116 available European observational sites, whereas 4 sites are qualified with NRA1. Meanwhile, both monthly results are relatively similar and in good agreement with the observations in the rest of regions. Regardless of small number of comparable sites, we examine that inter-annual δ¹⁸O variability for 15 years is also better reproduced with the ERA15 in Europe, while there is quite similar reproducibility in the rest of regions, too. This study thus reveals that inconsistency of temporal variations in precipitation between the two re-analyses mainly, but indirectly, causes this difference of isotopic reproducibility. The isotope analyses diagnose accuracy of two-dimensional water circulation fields in datasets with a particular focus on precipitation processes, but in a slightly different manner over middle and high-latitude and over low-latitude: severer over former regions.

In the fourth chapter, the author describes Iso-MATSIRO (Minimal Advanced Treatments of Surface Interaction and Runoff), a land surface model that includes stable water isotopes and simulates physically reasonable isotopic fluxes and reservoirs at the ground. The model calculates kinetic and equilibrium fractionation of HDO and H₂¹⁸O between ice, liquid, and vapor phases and separately considers soil surface evaporation, vegetation transpiration, evaporation from the canopy-intercepted reservoir, and snow sublimation. One-dimensional simulations with modeled meteorological forcings showed plausible features in the annual isotopic budget, seasonal variations of δ¹⁸O in soil moisture, diurnal variations of leaf water with some enrichment, and δ-diagram of representative surface reservoirs and fluxes. A subsequent, independent global simulation used Iso-MATSIRO coupled with an atmospheric isotope circulation model for a half year in 1998. Simulated precipitation δ¹⁸O was closer to observations than in a previous study, confirming the physical treatments of isotopes in the land surface processes.
The fifth chapter first implemented two different experiments to compare ICM developed by the author and Iso-AGCMs. One with constant field of evaporative isotopic ratio, and one with variable evaporative isotope fields which are the output from Iso-AGCM. The results showed, over the zonal regions around 30 to 60 degrees in both hemispheres, isotopes in evaporation have influences to the precipitation isotopes. Then, isotopically physical processes on the water surfaces are incorporated to the reanalysis-forced isotope circulation simulation. The 15-year simulation results show great improvement on the target regions, moreover, the systematic underestimation of precipitation isotopes over low latitudinal regions were also dissolved.

In the sixth chapter, the author tries to suggest some quantitative interpretation of precipitation isotopes with more precise and more reasonable global hydrologic cycles derived from the water circulation scheme that have been evaluated by using stable water isotopes in the previous chapters. In this chapter, the author investigated the dynamic motion of atmospheric water advection by an analytic method called colored moisture analysis (CMA), that allows for the estimation and visualization of atmospheric moisture advection from specific source regions. The CMA water transport model includes balance equations with the upstream scheme and, uses external meteorological forcings. The forcings were obtained from the GAME reanalysis, as the previous chapter. A numerical simulation with 79 global sections was run for April to October 1998. The results clearly showed seasonal variations in advection associated with large-scale circulation fields, particularly a difference between rainy and dry seasons associated with the Asian monsoon. The study also proposes a new definition of southwest Asian monsoon onset and decay, based on the amount of water originating from the Indian Ocean. Earliest onset occurs over southeastern Indochina around 16–25 May. Subsequent onset occurs in India one month later. These results agree with previous studies on the Asian monsoon onset. The CMA provides a clearer, more integrated view of temporal and spatial changes in atmospheric circulation fields, particularly Asian monsoon activities, than previous studies that focused only on one or two distinct circulation features, such as precipitation or wind speed. Furthermore, monsoon transition in a specific year, 1998, first became analyzable, whereas the previous studies used climatologies. Furthermore, the authors found the “precipitation age” is direct interpretation of precipitation isotopes over tropics, e.g., $\bar{\xi} = -0.25 \times \delta^{18}O + 5.8$ in Manaus, Brazil, is introduced.

The last chapter includes a summary and conclusions.
和文要旨

本論文は、水の安定同位体（重水素と重酸素を含む水分子：\( \text{HDO} \cdot \text{H}_{2}^{18}\text{O} \) を用いて地球の様々な水循環過程を解析することを目的としている。水の安定同位体を用いた研究は、水文学や生態学を中心とした様々な分野で古くから行われており、特に年代や起源、経路の特定及び更なる応用に有効なツールとして用いられてきた。中でも降水の同位体は、流出過程等の研究のはじめの入力値として必要であるが、アイスコアなどを用いた気候再構築等の研究を進めに際して現実気候の理解を深める必要があるという意味からも、非常に重要視されており、1960年代から全球規模の観測を中心として精力的に行われてきた。これまでに、非常に大きな時間的・空間的適合性が確認されるとともに、それらの変動に影響のある要素（例えば気象条件や地理条件・季節性）についていくつかの説明がなされてきた。しかし実際には、数多くの因子が相互に複雑に絡み合った影響を及ぼしているため、降水同位体比の時空間変動メカニズムの説明は定性的なものに終帯してきたとも言えぬではない。そこで本論文では、降水同位体比の変動メカニズムの定量的な解釈を可能にし、なぜこの雨がこの同位体を持つのか、この同位体比を持つ意味は何なのか、といった疑問への回答を試みた。降水同位体が大気中の循環過程や蒸発した際の地表面過程の影響を受けていることはすでに知られており、先ほどのような疑問に回答することは即ち、地球水循環過程を解明することに他ならない。

上記の目的を達成するため、著者は同位体全体輸送モデルを構築し、降水同位体比の実際の日単位変動から年々変動を相当程度再現することに成功した。また、再現性の良悪から用いた大気水循環場を評価する手法も併せて示した（2章と3章）。次に、地表面・水面における蒸発過程とそれに伴う同位体の物理挙動を合理的に表したモデルを用いて大気・陸面・水面結合シミュレーションを行い、再現性の良否を確認した（4章と5章）。最後に、蒸発時の空間情報や時間情報によって区別された（色付けされた）水蒸気が、大気中をどのように移動するかを詳細に解析し（色水解析）モンスーンの季節進行と降水起源変動の関係や、平均大気滞留時間と降水同位体比の関係について調査した（6章）に従い、各章に関する要旨を述べる。

第2章では、これまで合理的な説明がされていたなかった降水同位体比（\( ^{18}\text{O} \)）の日々変動に着目して、その変動要因の解明を行った。既往研究では、大気大循環モデルに同位体の物理過程を組み込んだIso-AGCM や、より簡略なレイン蒸留式を用いたモデルが用いられていたが、ここでは、両者の特徴を併せ持つレイリー型同位体循環モデルを新たに開発した。本モデルは、鉛直一層水平2次元のグリッドを持つ全球水蒸気輸送モデルであり、降水量・蒸発量・可降水量・鉛直積算水蒸気フラックスからなる水蒸気循環場を外力として大気水収支式に適用してグローバルな水蒸気輸送と混合過程を表現し、水蒸気から降水への同位体変分においても簡素なレイリー型を用いている。現実により近いGAME(全球エネルギーウェット循環実験観測アジアモンスーン領域プロジェクト)再解析データを循環場として用い、1998年4月から10月まで、1.25°×1.25°全領域の降水同位体比の時空間変動を推定したところ、タイの3地点において降水同位体比の日々変動を有意に再現し、全球の月単位の観測結果とも良く一致した。モデルによる降水同位体比の変動のうち、循環場の各要因別に分析すると、各単位の降水同位体の変動は水蒸気フラックスによる影響が一番大きく全球平均で47%（タ
イでは37%）だが、それは地域によって大きく異なった。また、より精度の良い衛星観測GPCP（Global Precipitation Climatology Project）を用いた場合に再現性が大きく向上した。

第3章は、第2章で開発した同位体循環モデルにNCEP/NCAR（アメリカ国立環境予測センター/国立中期予報センター）再解析（NRAI）とECMWF（ヨーロッパ中期予報センター）再解析（ERA15）を外力として与え1979年から1993年までの全球シミュレーションを行い、推定された降水同位体時空空間変動場をGNIP（Global Network in Isotopes in Precipitation）の観測同位体と月単位と年々変動の再現性について比較した。ERA15を用いた推定結果の方が、ヨーロッパを含む西〜中央ユーラシア大陆において再現性の精度が高く、当該地域における116の比較可能なサイトのうち87のサイトで有意な再現性が得られた。一方NRAIによる結果では4サイトのみであっ。ERA15の結果は、降水同位体比の年々変動も良く再現されており、月単位の結果と同じくヨーロッパ域により高い再現性が得られた。外力である気象4要素（降水・蒸発・可降水量・水蒸気フラックス）のうち、特に降水の精度の良し悪しが推定された降水同位体比変動の差異をもたらすが、その良し悪しの現れ方も地域によって異なり、高緯度地域での降水同位体比変動の再現は低緯度よりも困難であった。

第4章では、地表における氷とエネルギーのやり取りの結果として大気に戻される蒸発散フラックスの同位体比を合理的に求めるため、水の安定同位体の挙動を組み込んだ陸面モデル、Iso-MATSIRO（Minimal Advanced Treatments of Surface Interaction and Runoff）を構築した。モデルには、植物からの蒸散や土壌からの蒸発、樹冠による遮断水が蒸発、積雪の昇華・融解等に応じて、それぞれ異なる同位体物理が含まれている。3点の地域において1年間のシミュレーションを行い、同位体陆面モデルのパフォーマンスを調査したところ、同位体の年間収支、土壌水分同位体比の季節進行及び鉛直分布、植物水同位体比の日変化（昼間に重くなる）、土壌蒸発・蒸散・流出といったフラックス及び土壌水分・樹冠遮断・積雪といった貯留水のそれぞれのδ^{18}O−δD関係は、いずれも良好なものであった。最後にGAME再解析を用いて1998年の半年の大気−陸面結合シミュレーションを行ったところ、降水同位体比の推定値はより観測値に近づき、同位体陆面モデルで組み込まれた同位体物理の妥当性が確認された。

第5章ではまず、著者が開発した同位体循環モデル（ICM）と、前述の同位体を組み込んだ大気循環モデル（Iso-AGCM）との比較検証を行った。Iso-AGCMの外力である大気循環場をICMの外力とし、蒸発時の同位体プロセスを考慮する場合としない場合において、推定した降水同位体比の一致度をIso-AGCMと比較した。その結果、両半球の緯度30°〜60°において、蒸発フラックスの同位体比変動が降水同位体比に大きく影響していた。さらに、ICMに陸面モデル・水面モデルを組み込み、ERA15を用いた実験を行ったところ、中高緯度で季節変動がより正確に再現されると同時に、低緯度でもこれまで存在していた過小バイアスはほぼ消散し、降水同位体比の再現性は全球的に高まった。即ち、循環場の何を評価しているのかという質問に対して、第3章で得られた結論を修正して、「低緯度では主に降水・水蒸気フラックスによる蒸発気輸送過程を評価しているが、中高緯度では水蒸気輸送過程に加えて地面・水面過程も同位体で合わせて評価している」という結論が導き出された。

同位体を用いて評価された水蒸気輸送スキームを用いて、水循環の詳細を描き出し、降水同位体比との関係について再検討を加えたのが第6章である。ここでは、大気
中の水輸送の挙動を解明するために色水解析 (Colored Moisture Analysis: CMA) という手法を構築した。CMA とは、蒸発地域ごとに区別した水蒸気に対して大気水収支式を適用する全球 2 次元モデルに外部から循環場を与えることにより、水蒸気及び降水の起源を推定・視覚化する手法である。GAME 再解析を外力として 1998 年 4 月から 10 月までの計算を行ったところ、アジアモンスーン特有といえる、雨期と乾季で明瞭に異なる水蒸気の動きが視覚的に顕著に示された。また、アジアモンスーン域においてインド洋を起源とする水蒸気が増減する様子に着目し、その急激な増加をモンスーンの開始、減少を終了と定義してみると、一番早いモンスーンはインドシナ半島南部で 5 月 16 日頃に起こり、その後一、二ヶ月を経てインドに到達することがわかった。この結果は、雨量や風系、雲量等の気候値を使用したモンスーン開始 / 終了の定義と良く一致する。これまでは、雨量や風系といった一つの気象要素について注目しモンスーン等の挙動を分析していたが、CMA では、用いた 4 つの気象要素を複合的に考慮して起源別の水蒸気の輸送を計算しているため、総合的な大気循環場の時間空間変動がより明確に示されるようになった。またその結果は、気候値を用いないある特定の年のモンスーン進行を初めて示した。さらに、時間に対して水蒸気属性を与えられた「年間 CMA」を実行した。その結果、低緯度の特定の場所で降水同位体比の日々変動と「降水の年齢」のそれが非常に一致した地域が抽出され、ブラジルのマナウスでは、降水の平均年齢について \( \xi = -0.25 \times \delta^{18}O + 5.8 \) という関係が得られた。

最後に第 7 章にはこれまでの章のサマリー及び結論を示した。
Publication List

Chapter 2


Chapter 3


Chapter 4


Chapter 5

- Yoshimura, K. and K. Sturm, Practical interpretation of isotopic information in precipitation, manuscript in preparation.
Chapter 6


Appendix A

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Chapter 1

Introduction

Stable water isotopes (HDO and H$_2^{18}$O) have been used in many multi-disciplinary studies of hydrologic cycles on various scales, and have provided large number of new insights into the age, origin, and pathway of water movement [Kendall and McDonnell, 1998]. For example, isotope hydrograph separations on hillslope to catchment scales have changed the classical treatments on the runoff processes [McDonnell et al., 1991], and laboratory-based knowledge on isotopic fractionations with intensity of evaporation were applied on the biological studies with evapotranspiration separations [e.g., Dawson, 1993]. Likewise, studies on isotopes in precipitation have started in 1960’s, and have shown great number of new insights, such as the meteoric water line [Craig, 1961] or many spatio-temporal isotopic effects [e.g., Clark and Fritz, 1997].

The studies on isotopes in precipitation have been trying to understand the formation processes that make the isotopic spatio-temporal heterogeneity, and they are regarded very important not only because many isotopic studies on ground hydrology must start with them, but also because they are used as proxy for reproduction of paleo-climate. Currently, we all know the variations of precipitation isotopes are derived from combinations of meteorological status, temporal and geographical conditions, and so on. The statement is most likely true, but because there are too many explanation variables and vague relationships with precipitation isotopes or/and inbetween, the processes have not been quantitatively explained, so that the deterministic mechanism, such as “why this rain has these isotopic values” or “what these isotopic values tell us”, remain unclear. This thesis challenges to answer these questions.

The author first creates models that incorporate isotopic processes to quantitatively induce the mechanism of observed variability of precipitation isotopes. The model is designed to adopt additional physical processes as model components that possibly influence the variability, so that author can examine degree of the influences to precipitation isotopes. Concurrently, the author suggests a new use of water isotopes, namely “evaluation of the global hydrologic cycles by precipitation isotopes”. The framework of the evaluation is developed in this thesis. The knowledge from the inductive modeling study above improves the performance of the evaluation, and provides characteristic and interpretation of evaluated results. Furthermore, analytic investigations of the dynamic motion of atmospheric water advection are derivatively performed for finding of more direct interpretation of the isotopology.

Each of six chapters after this chapter individually or successively implements and investigates
the topic(s) or part of the topic(s) above. A summary and conclusions of this thesis then follows. The structure of the thesis is described in Figure 1.1. Brief descriptions of each chapter are shown below.

![Diagram](image)

Figure 1.1: Schematic representation of the global hydrologic cycles and corresponding chapters in this thesis

The second chapter seeks to explain the causes of short-term (1–10 days) variability in precipitation isotopes, on which two types of models, *i.e.*, Rayleigh-type models and isotope-AGCMs (Atmospheric General Circulation Models), in the past have not been able to examine clearly. A new water isotope circulation model on a global scale that includes a Rayleigh equation and the use of external meteorological forcings is developed. Transport and mixing processes of water masses and isotopes that have been neglected in earlier Rayleigh models are included in the new model. A simulation of $^{18}$O for 1998 is forced with data from the Global Energy and Water Cycle Experiment (GEWEX) Asian Monsoon Experiments (GAME) reanalysis. The results are validated by GNIP (Global network of isotopes in precipitation) monthly observations with correlation *R* = 0.76 and a significance level > 99%, and daily observations at three sites in Thailand with similar correlation and significance. A quantitative analysis of the results shows that among three factors that cause isotopic variability, the contribution of moisture flux is the largest, accounting for 37% at Chiangmai, and 46% globally. This highlights the importance of transport and mixing of airmasses with different isotopic concentrations. A sensitivity analysis of the temporal and spatial resolution required for each variable is also made, and the model is applied to two additional datasets. The more accurate GPCP (Global Precipitation Climatology Project) precipitation dataset yields improved model results at all three observation sites in Thailand. The National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis allows the simulation to cover two years, reproducing reasonable interannual isotopic variability.
1. INTRODUCTION

A global Rayleigh-type \( H_2^{18}O \) circulation model developed in the second chapter was forced by two reanalysis products, NCEP/NCAR reanalysis (NRA1) and the European Centre for Medium-Range Weather Forecasts (ECMWF) 15-year reanalysis (ERA15) datasets, and the GNIP database was used to validate for 1979-1993 on monthly to inter-annual scales, in the third chapter. We find out that the model monthly reproducibility is better with ERA15, particularly in Europe; the number of sites with a statistically >99% significant correlation is 87 out of 116 available European observational sites, whereas 4 sites are qualified with NRA1. Meanwhile, both monthly results are relatively similar and in good agreement with the observations in the rest of regions. Regardless of small number of comparable sites, we examine that inter-annual \( \delta^{18}O \) variability for 15 years is also better reproduced with the ERA15 in Europe, while there is quite similar reproducibility in the rest of regions, too. This study thus reveals that inconsistency of temporal variations in precipitation between the two re-analyses mainly, but indirectly, causes this difference of isotopic reproducibility. The isotope analyses diagnose accuracy of two-dimensional water circulation fields in datasets with a particular focus on precipitation processes, but in a slightly different manner over middle and high-latitude and over low-latitude: severer over former regions.

In the fourth chapter, the author describes Iso-MATSIRO, a land surface model that includes stable water isotopes and simulates physically reasonable isotopic fluxes and reservoirs at the ground. The model calculates kinetic and equilibrium fractionation of HDO and \( H_2^{18}O \) between ice, liquid, and vapor phases and separately considers soil surface evaporation, vegetation transpiration, evaporation from the canopy-intercepted reservoir, and snow sublimation. One-dimensional simulations with modeled meteorological forcings showed plausible features in the annual isotopic budget, seasonal variations of \( \delta^{18}O \) in soil moisture, diurnal variations of leaf water with some enrichment, and \( \delta \)-diagram of representative surface reservoirs and fluxes. A subsequent, independent global simulation used Iso-MATSIRO coupled with an atmospheric isotope circulation model for a half year in 1998. Simulated precipitation \( \delta^{18}O \) was closer to observations than in a previous study, confirming the physical treatments of isotopes in the land surface processes.

The fifth chapter first implemented two different experiments to compare ICM developed by the author and Iso-AGCMs. One with constant field of evaporative isotopic ratio, and one with variable evaporative isotope fields which are the output from Iso-AGCM. The results showed, over the zonal regions around 30 to 60 degrees in both hemispheres, isotopes in evaporation have influences to the precipitation isotopes. Then, isotopically physical processes on the water surfaces are incorporated to the reanalysis-forced isotope circulation simulation. The 15-year simulation results show great improvement on the target regions, moreover, the systematic underestimation of precipitation isotopes over low latitudinal regions were also dissolved.

In the sixth chapter, the author tries to suggest some quantitative interpretation of precipitation isotopes with more precise and more reasonable global hydrologic cycles derived from the water circulation scheme that have been evaluated by using stable water isotopes in the previous chapters. In this chapter, the author investigated the dynamic motion of atmospheric water advection by an analytic method called colored moisture analysis (CMA), that allows for the estimation and visualization of atmospheric moisture advection from specific source regions. The CMA water transport model includes balance equations with the upstream scheme
and, uses external meteorological forcings. The forcings were obtained from the GAME reanalysis, as the previous chapter. A numerical simulation with 79 global sections was run for April to October 1998. The results clearly showed seasonal variations in advection associated with large-scale circulation fields, particularly a difference between rainy and dry seasons associated with the Asian monsoon. The study also proposes a new definition of southwest Asian monsoon onset and decay, based on the amount of water originating from the Indian Ocean. Earliest onset occurs over southeastern Indochina around 16–25 May. Subsequent onset occurs in India one month later. These results agree with previous studies on the Asian monsoon onset/ end. The CMA provides a clearer, more integrated view of temporal and spatial changes in atmospheric circulation fields, particularly Asian monsoon activities, than previous studies that focused only on one or two distinct circulation features, such as precipitation or wind speed. Furthermore, monsoon transition in a specific year, 1998, first became analyzable, whereas the previous studies used climatologies. Furthermore, the authors found the “precipitation age” is direct interpretation of precipitation isotopes over tropics, \( \xi = -0.25 \times \delta^{18}\text{O} + 5.8 \) is introduced in Manaus, Brazil.
Chapter 2

A Quantitative Analysis of Short-term $^{18}$O Variability with a Rayleigh-type Isotope Circulation Model

Abstract:
Stable water isotopes (D and $^{18}$O) in precipitation have large spatial and temporal variability and are used widely to trace the global hydrologic cycle. The two models that have been used in the past to examine the variability of precipitation isotopes are Rayleigh-type models and isotope-AGCMs. The causes of short-term (1–10 days) variability in precipitation isotopes, however, remain unclear. This study seeks to explain isotope variability quantitatively at such scale. A new water isotope circulation model on a global scale that includes a Rayleigh equation and the use of external meteorological forcings is developed. Transport and mixing processes of water masses and isotopes that have been neglected in earlier Rayleigh models are included in the new model. A simulation of $^{18}$O for 1998 is forced with data from the GAME reanalysis. The results are validated by GNIP monthly observations with correlation $R=0.76$ and a significance level $>99\%$, and daily observations at three sites in Thailand with similar correlation and significance. A quantitative analysis of the results shows that among three factors that cause isotopic variability, the contribution of moisture flux is the largest, accounting for 37\% at Chiangmai, and 46\% globally. This highlights the importance of transport and mixing of air-masses with different isotopic concentrations. A sensitivity analysis of the temporal and spatial resolution required for each variable is also made, and the model is applied to two additional datasets. The more accurate GPCP precipitation dataset yields improved model results at all three observation sites in Thailand. The NCEP/NCAR reanalysis allows the simulation to cover two years, reproducing reasonable interannual isotopic variability.
2.1 Introduction

Stable isotopes of water, D and $^{18}$O, have been used as tracers of the hydrologic cycle for more than 50 years. The first studies that considered isotopes focused on isotopic enrichment and the isotopic uniformity of sea water [Gilletian, 1934; Epstein and Mayeda, 1953; Friedman, 1953]. Dansgaard [1964] made the first successful trial to quantitatively understand the water isotopes in precipitation. Subsequent studies used isotopic characteristics to analyze precipitation at different spatial and temporal scales, as Welker [2000] classified. The studies have revealed great heterogeneity in the temporal and spatial distributions of precipitation isotopes. These precipitation characteristics are well known as several isotopic effects [e.g., Clark and Fritz, 1997].

Past studies frequently used Rayleigh’s distillation process (the Rayleigh equation as described below) and developed models [e.g., Dansgaard, 1964; Jouzel and Merlivat, 1984; Ciais and Jouzel, 1994] to help explain spatial and temporal variability in precipitation isotopes. The physical basis behind the Rayleigh equation that governs precipitation isotope is that isotopes are exchanged between moisture in rising air and falling droplets and precipitation essentially ‘forgets’ any isotopic label from in-cloud processes as isotopic equilibrium with the ambient air is established [Friedman et al., 1962; Gat, 2000].

A common weakness in Rayleigh models is oversimplification of the full complexity of the global hydrological cycle [Hoffmann et al., 2000]. In particular, Rayleigh models were assumed unable to account for the mixing of air masses of different origin and the high variability of vapor pathways from the site of evaporation to the site where condensation takes place and finally the water precipitates. Consequently, Rayleigh equations have been used only for qualitative explanations of observed isotopic characteristics. For example, isotopic depletion of precipitation far from the ocean or at high altitudes.

Therefore, it was a big step when Joussaume et al. [1984] incorporated the physics of water isotopes into an atmospheric general circulation model (AGCM), the model of the Laboratoire de Meteorologie Dynamique (LMD). Subsequently, Jouzel et al. [1987] embedded the water isotopes in the GISS (Goddard Institute for Space Studies) model, Hoffmann et al. [1998] in the ECHAM (European Center model modified in Hamburg) model, Mathieu et al. [2002] in the GENESIS (Global Environmental and Ecological Simulation of Interactive Systems) model, and Noone and Simmonds [2002] in the MUGCM (Melbourne University GCM). Thereafter, recent similar studies have been trying to couple stable isotopes with mesoscale models inquiring more precise isotopic phenomena such as internal structure of storms [e.g., Gedzelman et al., 2003; Lawrence and Gedzelman, 2003]. AGCMs already takes account of the complexity of the atmospheric processes leading to the global hydrological cycle, which is oversimplified in Rayleigh-type models. AGCMs that include isotopes should be better at reproducing observed spatial and temporal heterogeneity of precipitation isotopes. Indeed, isotope AGCM studies have reasonably reproduced the major isotopic features in precipitation, and have revealed the large-scale characteristic of observed isotope-climate relationships [Mathieu et al., 2002].

Such reproductions, however, have been reasonable only for monthly or annual averages. Many regional isotopic studies have shown that short-term isotopic variability in precipitation is much greater than that over a season or a month. Such studies include Jacob and Sonntag
[1991] for western Europe, Njitchoua et al. [1999] for Africa, Rindsberger et al. [1983] for Israel, Welker [2000] for USA, Shimada et al. [1998] for Japan, and the present study for Thailand, as detailed below. As Hoffmann et al. [2000] noted, isotope AGCM studies cannot reproduce this short-term isotopic variability, perhaps because of spatial resolution limitations. They noted that rain formation processes that influence isotopic signals of single precipitation events occur on small spatial scales, about 30 km in the horizontal [Ehnhall and Östlund, 1970], and current models, with horizontal resolutions of more than 100 km, cannot accurately account for such processes.

The failure of isotope-AGCMs to reproduce short-term variability may have other causes, however. Isotopic variability with time has been observed at spatial scales of greater than 100 km in the horizontal, despite rain formation processes on smaller horizontal scales. For example, Yamanaka et al. [2002] showed that spatial variability of the temporal variability at 18 observation sites in a 150 km by 150 km region around Tokyo, Japan, was small compared to event-based temporal variability. In addition, daily precipitation isotopic data have been collected and analyzed at three sites in Thailand (Bangkok, Sukhothai, and Chiangmai; see Figure 2.1) since 1998. These data show similar temporal variability (Figure 2.2), even though Bangkok and Chiangmai are 600 km apart and single precipitation events on small spatial scales are common during the rainy season in Thailand. This spatially uniform temporal variability suggests that large scale moisture transport dominates the control of short-term isotopic variability. Furthermore, regional processes at smaller scales, such as the formation of cloud liquid and solids in single storms and their convective redistribution within the clouds, are of secondary importance. If so, the cause of the failure is likely due to that short-term-averaged meteorological field from AGCMs is less realistic than monthly- or annual-averaged values. Short-term isotopic variability hence has not been accurately reproduced.

This study will quantitatively explain the short-term variability of precipitation isotopes, focusing on whether moisture transport at large scales is important. A new model is developed that simulates the global circulation of atmospheric water and its isotope. Unlike isotope-AGCMs, the new model incorporates external meteorological datasets that are more realistic than the internally generated circulation fields in the AGCMs. The more realistic external datasets are taken from reanalyses that include long-term gridded global datasets of key variables constrained and assimilated by observations [Bengtsson and Shukla, 1988]. It is difficult, however, to incorporates all atmospheric physics including in-cloud microphysical processes or convection activities, because of the limited number of variables included in the reanalyses. It is hence useful to include a Rayleigh equation to simulate isotopic fractionation in the precipitation. The Rayleigh equation, as noted above, describes an isotopic equilibrium relationship between precipitation and ambient air, regardless of any isotopic variability introduced by in-cloud processes. Thus, the Rayleigh model, which includes previously neglected transport and mixing of water masses and their isotopes, is also suitable for evaluating the impact of transport at large scale on precipitation isotope. The new model is therefore named a Rayleigh-type isotope circulation model that is intermediate between Rayleigh-type models and isotope-AGCMs.

The second section briefly introduces the fundamentals of the Rayleigh equation and of stable isotopes in water. The Rayleigh-type isotope circulation model is detailed in the third section. This is followed in the fourth section by a description of, and results from, a control simulation.
A quantitative analysis of short-term isotopic variability is also demonstrated. The fifth section includes sensitivity experiments and simulations using different meteorological datasets. The last section contains a summary and conclusion.

![Map of Thailand showing Chiangmai, Sukhothai, and Bangkok](image)

**Figure 2.1:** Location of the observation sites, Bangkok, Sukhothai, and Chiangmai, in Thailand.

### 2.2 Fundamentals of Water Isotopes

#### 2.2.1 Terminology

The stable water isotopes D and $^{18}$O are measured in units of parts per thousand ($\%_o$) relative to a standard (SMOW: standard mean ocean water) composition. For example, $\delta^{18}$O values are calculated by

$$
\delta^{18}O = ((R_A/R_S) - 1) \times 1000 [\%_o]
$$

(2.1)

where $R_A$ and $R_S$ denote the ratios in the sample and standard, respectively, of the heavy to light isotope ($^{18}$O/$^{16}$O).

#### 2.2.2 Fractionation and Fractionation Factor

Because of mass differences, H and D and $^{16}$O and $^{18}$O have different chemical and physical properties. Such differences are manifested as a fractionation effect. When water changes phase from liquid to gas, the heavy isotopes preferentially enrich in the liquid. The fractionation is
Figure 2.2: Observed $\delta^{18}O$ in precipitation at the three sites in Thailand (a) in 1998 and (b) in 1999.
expressed as the isotope fractionation factor $\alpha$

$$\alpha_{A,B} = \frac{R_B}{R_A}$$  \hspace{1cm} (2.2)

where $R_A$ and $R_B$ are the isotope ratios of phase A and phase B, respectively. When $\alpha_{A,B} > 1$, the heavy isotopes are enriched in phase B. When $\alpha_{A,B} < 1$, the heavy isotopes are depleted in phase B.

### 2.2.3 Rayleigh Equation

As water changes phase, the evolution of the isotopic composition is described by a Rayleigh equation with fractionation factor $\alpha$ as follows for $^{18}$O,

$$\frac{R}{R_0} = \frac{1 + 10^{-3\delta^{18}O}}{1 + 10^{-3\delta^{18}O}} = \left( \frac{W}{W_0} \right)^{\alpha-1}$$  \hspace{1cm} (2.3)

where $R_0$ and $R$ are the initial and final isotope ratios, $\delta^{18}$O and $\delta^{18}$O are the respective $\delta$-values, and $W_0$ and $W$ are the initial and final amounts of water in the final phase.

### 2.3 Model Description

The isotope circulation model used in this study is simpler than those of previous isotope-AGCM studies. The model consists of a global grid ($1.25^\circ \times 1.25^\circ$ or $2.5^\circ \times 2.5^\circ$ and one vertical layer), and each grid is approximated as a trapezoid. As illustrated in Figure 2.3, the model computes water transport by maintaining the atmospheric water balance [Oki et al., 1995] with the upstream scheme in each grid at every time step. The model time step ($\Delta t$) is 10 minutes. This scheme, however, often does not satisfy CFL (Courant-Friedrichs-Lewy) condition near the Poles because of the short distance between grid points there. Calculations are therefore not performed poleward of $85^\circ$N or $85^\circ$S, and a constant value ($-30\%$) for the isotope composition of vapor is assigned there. This abbreviation may cause uncertainties in the model at high latitudes, but isotopic influence of those polar regions on main target areas of the present study, the tropics and the subtropics including Thailand, is assumed as negligible. Finally, during every time step in water transport calculation, the isotopic compositions ($\delta$-value) of both vapor and precipitation are calculated using isotopic mass-balance and Rayleigh fractionation as described below.

One notable feature of the model is its single layer in the vertical. Vertical profiles of water and isotope composition are included as vertical integrations. In this way, processes influence the local isotopic composition of precipitation, including the formation of cloud liquid or solid and their redistribution by convection [e.g., Hoffmann et al., 2000], all cancel out. Indeed, the model considers only the fractionation generated by condensation at large horizontal scales using the fractionation factor $\alpha$. Here, 'large scale condensation' means the spatially averaged accumulating precipitation in each grid.

There is one more distinctive model characteristic. When the model calculates the atmospheric water balance, it uses forcing variables computed from external meteorological datasets such as reanalyses or observations. Thus, the model can reproduce the water cycle more realistically than isotope-AGCM studies.
2. RAYLEIGH-TYPE $^{18}$O CIRCULATION MODEL

![Diagram of ICM processes]

Figure 2.3: Schematic representation of ICM processes in each time step.

2.3.1 Atmospheric Water Balance

Oki et al. [1995] used the atmospheric water balance equation

$$\frac{\partial W}{\partial t} = - \nabla \cdot \bar{Q} + (E - P) \quad (2.4)$$

where $W$, $- \nabla \cdot \bar{Q}$, $P$, and $E$ represent precipitable water, horizontal water vapor flux convergence, precipitation and evaporation, respectively. $\bar{Q}$ is the vertically integrated vapor flux vector; its components are the zonal and the meridional fluxes, $Q_\lambda$ and $Q_\phi$, respectively. $Q_\lambda$, $Q_\phi$, and $W$ are defined as follows:

$$Q_\lambda \equiv \int_0^{p_0} q u \frac{dp}{g} \quad (2.5)$$

$$Q_\phi \equiv \int_0^{p_0} q v \frac{dp}{g} \quad (2.6)$$

$$W \equiv \int_0^{p_0} q \frac{dp}{g} \quad (2.7)$$

where $q$, $u$, $v$, $g$, $p$, and $p_0$ are specific humidity, zonal wind speed, meridional wind speed, gravitational acceleration, atmospheric pressure, and surface pressure.

Here, we tried a test calculation using $- \nabla \cdot \bar{Q}$, $P$, and $E$ from the GAME (GEWEX Asian Monsoon Experiments) reanalysis [Yamazaki et al., 2001]. Figure 2.4 shows the results and the monthly change in precipitable water is unrealistically large, up to 150 mm per month. Such problematic results have occurred in similar atmospheric water balance studies [e.g., Bosilovich and Schubert, 2001; Kanae et al., 2001] because the balance equation is not always closed. In fact, if external meteorological data are used, the unclosed nature of the balance equation is inevitable. The water balance equation (2.4) is discretized

$$W_{(t+\Delta t)} = W_{(t)} - \nabla \cdot \bar{Q} \Delta t + (E - P) \Delta t \quad (2.8)$$

Regarding that the water balance is not closed, calculated $W_{(t+\Delta t)}$ is used for following isotopic balance equations below. In the next time step, however, precipitable water on the right side of (2.8) is taken from observed (or reanalyzed) precipitable water instead of using $W_{(t+\Delta t)}$ in the previous time step.
Figure 2.4: Global distribution of precipitable water divergence for August 1998 calculated by only moisture convergence, precipitation and evaporation from the GAME reanalysis. Solid (dashed) contours denote positive (negative) divergence, and the area with diverged values over 50 mm/month (below −50 mm/month) are shaded by light (dark) gray.
2.3.2 Isotope Mass-Balance Coupled with Rayleigh Fractionation

Discretized water balance equation (2.8) is combined with isotopic mass-balance to yield

$$
\delta w(t+\Delta t)W(t+\Delta t) = \delta w(t)W(t) - \nabla \cdot \delta w\bar{Q}\Delta t + \delta eE\Delta t - \delta pP\Delta t
$$

(2.9)

where $\delta w$, $\delta e$, and $\delta p$ are the isotope ratios of precipitable water, evaporation, and precipitation. The term $\nabla \cdot \delta w\bar{Q}$ is defined as

$$
\nabla \cdot \delta w\bar{Q} = \frac{1}{R_e \cos \phi} \left( \frac{\partial \delta wQ_\lambda}{\partial \lambda} + \frac{\partial \delta wQ_\phi \cos \phi}{\partial \phi} \right)
$$

(2.10)

where $R_e$ is radius of the Earth, and $\lambda$ and $\phi$ denote longitude and latitude, respectively. If only precipitation causes isotope fractionation, the Rayleigh equation (2.3) can be rewritten as

$$
\delta w = \left( \frac{W}{W^*} \right)^{\alpha^{-1}} (1 + 10^{-3}\delta w^*) \times 10^{3}
$$

(2.11)

where $W$ and $\delta w$ are the precipitable water and its isotopic ratio, respectively; $W^*$ and $\delta w^*$ are those just before precipitation occurs ($W^* = W + P\Delta t$); and $\alpha$ is the fractionation factor. Then, if mass and isotope balance are conserved, the isotope ratio of the precipitation is

$$
\delta wW = \delta w^*W^* - \delta pP\Delta t
$$

$$
\delta p = \frac{\delta w^* - \delta w f}{1 - f}
$$

(2.12)

where $f = W/W^*$. Finally, the weighted daily mean precipitation isotope ratio is calculated.

2.4 Control Simulation

2.4.1 Description

The model domain for the control simulation is 1.25°×1.25° and one vertical layer. Simulation results will be validated by monthly and daily observations. However, available daily data are only in Thailand for a moment that are sampled at three sites under a part of activities of GEWEX Asian Monsoon Experiment (GAME). The focus of this study is therefore an area influenced by the Asian Monsoon, specifically Thailand. A reanalyzed dataset that is achieved also by the GAME activities, called the GAME reanalysis Ver.1.5, is used for all forcing variables, such as $W$, $\bar{Q}$, $E$, $P$. The GAME reanalysis has been constrained by many assimilated observations in the Asian Monsoon region. Moreover, the dataset resolution is finer in temporal (6-hourly) and spatial (1.25°×1.25°) than other reanalyses. The use of the GAME reanalysis is therefore reasonable for the control simulation, despite the short 7-month period reanalyzed (0UTC 1 April to 18UTC 31 October 1998).

This study focuses on the short-term variability of $^{18}$O for the present and the circulation of $^{18}$O will be simulated. The fractionation factor $\alpha$ is set to 1.0094 at 25°C [Majoube, 1970]. Further, to help determine the $^{18}$O value of the evaporating vapor, $\delta e$, the global surface is
divided into three parts: seas, land between 40°N and 40°S, and everywhere else. The values for the three regions are -9.4 ‰, -10 ‰ and -15 ‰, respectively. These values are grounded on assumptions as follows, respectively: isotopically uniform (0 ‰) and limitless seawater is assumed to make uniform -9.4 ‰ evaporating vapor by equilibrium fractionation at 25 °C (α = 1.0094) over all seas; regarding the weighted zonal averages of observed annual mean δ¹⁸O in precipitation between 40°N and 40°S are quite uniform, approximately -5 ‰ [Mathieu et al., 2002], a composite of half transpiration from plants (with no isotopic fractionation [Zimmerman et al., 1967]) and half evaporation from bare soil and watery surface (with fractionation at normal temperature) makes -10 ‰ for δe in 40°N–40°S; and more isotopic depletion of the weighted mean precipitation isotopes for the rest of regions (>40°N and >40°S), approximately -10 ‰, makes -15 ‰ for δe by the same assumption as above. Note that isotopic behavior that accounts of interactions between land surface and atmosphere is still uncertain because of model complexity and lack of observational validity [Mathieu and Bariac, 1996], so that the constant values in time series and the rough spatial division into two parts are made for δe on lands.

Specifications for the control simulation are displayed in Table 2.1. An initial δ-value of atmospheric water δₒ₀ is set to 0 ‰ at all grid points. The model runs for the first month (April 1998) to reach an isotopic steady state, and re-runs from 1 April with the δ-value on 30 April in the first run at each grid point.

### 2.4.2 Results

**Comparison with Monthly Global Observations**

At first, monthly and global simulation results are validated. Global maps of δ¹⁸O in monthly precipitation are shown in Figure 2.5. The 'latitude effect' and 'continental effect' (e.g., Ingraham [1998]) are clearly reproduced on a global scale. Seasonality is also evident. For example,
differences between April and August are significant, especially the depletion in the northern sub-tropics and the enrichment near the Equator. Further, we can see isotopic depleted bands near the Equator on all figures, and they probably correspond with ITCZ and its seasonal shift.

Figure 2.6 shows a comparison in August with monthly GNIP (Global Network of Isotopes in Precipitation) [IAEA/WMO, 2001] observations. The GNIP data are weighted monthly means of precipitation for several years, but interannual variability is not as large as seasonal or daily variability, so these values can be used for comparison.

Figure 2.6(a) displays the success of the model in reproducing features in the oceans, northern Eurasia, South America, Australia, Africa, and Southeast Asia, within 6 % error range. Figure 2.6(b) shows that the model successfully reproduced spatial characteristics on a global scale. The correlation coefficient was 0.76, with a significance level exceeding 99%, and the root mean square error (RMSE) was 4.80 %.

A quantitative comparison of these values with earlier similar studies is quite difficult because the previous isotope-AGCMs emphasized the relationship between local temperature and isotope. However, some of the earlier studies include error contour maps as in Figure 2.6(a), and according to those maps, their one-to-one results seem less accurate than their temperature-isotope results. For example, Mathieu et al. [2002] has errors as large as 14% for mountainous regions such as Tibet or the Rocky Mountains. In addition, a recent isotope-AGCM study by Vuille et al. [2003] includes the results of different isotope-AGCMs and several one-to-one comparisons for the weighted annual means in Americas. The reported correlation coefficients are in the range between 0.44 to 0.79. Thus the present model results are accurate enough for a first order global estimate. However, isotope ratios are sometimes smaller than observed values. These underestimates suggest that the model needs to be improved.

The GNIP dataset for this comparison includes only 389 observation points. The observation network is therefore not dense enough to describe global precipitation isotope distributions accurately. Furthermore, the minimum temporal unit in the GNIP dataset is as long as a month, but monthly observation data are not always available. Observations with finer spatial and temporal scales, such as the observations for Thailand shown above, are required for global validation.

**Comparison with Daily Observations in Thailand**

The previous discussion highlighted the good agreement between global and monthly 18O simulations and observations. Figure 2.7 shows the simulated results and a comparison with daily observations at three sites in Thailand, which indicates that the model reproduced the variability of the precipitation isotopes at both short and seasonal time scales. In detail at Chiangmai, a depleted peak on near 1 July and a followed enrichment until 6 July and a depletion again until 11 July were well reproduced. A continuous increasing trend from the end of July to the mid of August and a sudden depletion around 20 August were also traced. The correlation coefficients and RMSEs are 0.76 and 4.23% at Chiangmai (N18.8°, E99.0°), 0.72 and 4.10% at Sukhothai (N17.0°, E99.8°), and 0.56 and 3.50% at Bangkok (N13.8°, E100.5°). All correlation coefficients are significant at a level exceeding 99%.
Figure 2.5: Distribution of predicted $\delta^{18}O$ values in the monthly precipitation from the control simulation. (a)-(g) correspond to each month from April to October, 1998. Contour interval is 3‰.
Figure 2.6: Global comparison, using monthly data, between the simulation predictions and GNIP observations (shown is Aug. 1998): (a) Geographical distribution map. Contour interval is 3 ‰. Contours with negative values are drawn by dashed lines. The regions with error exceeding 6 ‰ (below −6‰) are shaded by light (dark) gray. (b) Scatterplot diagram of each observation sites. The correlation coefficient is 0.76 and RMSE is 4.80 ‰.

On the other hand of the well reproduction of the isotopic variability, however, the discrepancy of over-depleted values in the simulation by 4 ‰ is apparent as same as in the global monthly results (previous section). This may be due to several uncertainty in the model (discussed in the last paragraph of section 2.6) and the errors included in the external variables, the GAME reanalysis, to some extent (discussed in the next section). It is though meaningful that however much those uncertainty and errors are included, the good reproduction of the short-term variability are made by the simple Rayleigh-type circulation model.

To explain the control factors of isotopic variability, we will focus on the results for Chiangmai. Figure 2.8 shows the temporal fluctuations of δ¹⁸O in vapor and precipitation, and the forced precipitation. There is a little negative correlation between precipitation and its isotopic composition ($R = -0.34$). On the other hand, there is a very strong correspondence between the isotopes in the vapor and the precipitation ($\delta p \approx \delta w + 9$, $R = 0.99$).

This indicates that the impact of the amount of precipitation at Chiangmai, i.e., “amount effect”, is insufficient to explain the short-term isotopic variability in the precipitation. On the other hand, the isotopic composition of the precipitation is strongly dependent on the isotopic composition of the vapor above the site. Obviously, this close relationship is due to a link between water vapor isotopes and precipitation isotopes through the fractionation factor. It tells, however, that what controls the water vapor isotopes also controls the variability of the precipitation isotopes. The present model calculates isotopic composition of water vapor by taking into accounts of horizontal water vapor transport that is influenced by precipitation and evaporation along the transport trajectory. In other words, water and isotopic mass balance on over 100 km horizontal scale control variability of water vapor isotopes. The good agreements between the model results and the daily and monthly observations therefore imply that moisture
transport at large scales (>100 km) dominantly controls the short-term isotopic variability of precipitation, particularly over Thailand. The following section provides a more precise account of the control factors of the short-term variability.

![Graphs showing results for Chiangmai, Sukhothai, and Bangkok.](image)

Figure 2.7: Simulation results compared with daily observations at (a) Chiangmai, (b) Sukhothai and (c) Bangkok.
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Figure 2.8: Time series variability of $\delta^{18}$O in vapor (black-solid) and precipitation (gray-dashed) and rainfall amount (bars) at Chiangmai.

### 2.4.3 Control Factors of Short-term Isotopic Variability

The control factors of short-term vapor isotopic variability will be quantitatively analyzed in this section. This model includes only four forcing variables, $\bar{Q}$, $E$, $W$, and $P$, to simulate atmospheric water circulation. Of the four, the change of precipitable water is dependent on other forcings, as shown in discretized water balance equation (2.8). Therefore, precipitable water has little impact on short-term isotopic variability. Furthermore, because isotope-related parameters such as $\alpha$ and $\delta e$ are constant for the period in the control simulation, the three forcings $\bar{Q}$, $E$, and $P$ must govern the isotopic circulation and generate spatial and temporal variability in the water vapor isotopes.

From the isotope balance equation (3.2), the change in the isotopic ratio in the precipitable water $\Delta \delta w$ consists of three terms by each derivation

$$\Delta \delta w = (\Delta \delta w)_Q + (\Delta \delta w)_E + (\Delta \delta w)_P$$

(2.13)

If the sizes of these terms are calculated, the impacts of each forcing on the isotopic variability can be quantitatively examined. Results for the control simulation are given in Figure 2.9. Moisture flux both depletes and enriches the isotopes, but precipitation only depletes and evaporation only enriches. More significantly, the fluctuations in $(\Delta \delta w)_Q$ and $(\Delta \delta w)_P$ are very large relative to the nearly constant $(\Delta \delta w)_E$. Thus, it is likely precipitation or moisture flux that generates any sudden isotopic depletion or enrichment in precipitation isotope variability. The daily average of the absolute value is 1.02%/oo for $(\Delta \delta w)_Q$, 0.73%/oo for $(\Delta \delta w)_E$, and 1.01%/oo for $(\Delta \delta w)_P$. In Chiangmai for the simulation period, isotopic variability is controlled to some degree by moisture flux and precipitation, and to a lesser degree by evaporation. The averaged contributions to the isotopic variability are 37% for moisture flux, 37% for precipitation, and 26% for evaporation.

Figure 2.10 shows the global distribution of each contribution to the isotopic variability. The
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figures show clear geographical distinctions. Moisture flux has dominant role in arid regions, precipitation dominates in rainy regions, and evaporation is important over warm maritime regions. Over the entire globe, the contribution of moisture flux is the largest. The means weighted by the amount of precipitable water are 46.3% for moisture flux, 23.2% for evaporation, and 30.5% for precipitation. The simple arithmetic means are 47.3%, 23.5%, and 29.2%, respectively.

In this analysis, the contribution of precipitation at the site was only evaluated. In other words, the impact of precipitation at the surrounding grids was not considered. However, Equation (2.10) shows that the impact of moisture flux on isotope circulation is dependent on the spatial isotopic concentration gradient $(\nabla \delta w)$. The concentration gradient is generated by precipitation, because only precipitation causes fractionation in this model. Thus, if the effect of surrounding grid points is accounted for, it becomes difficult to separate the distinctive impacts of moisture flux and precipitation on the isotopic concentration gradient. Nevertheless there may be uncertainties as raised above, the quantitative findings discussed here concerning the control factors of short-term isotopic variability are meaningful, because there has been no quantitative explanation for the control factors in previous studies.

![Figure 2.9: Time series variability of the three factors forcing daily changes of $\delta w$ at Chiangmai: $(\Delta \delta w)_Q$, $(\Delta \delta w)_E$, and $(\Delta \delta w)_P$ are shown as solid, dashed, and dotted lines, respectively.](image)

**2.5 Simulations Using Different Datasets**

**2.5.1 Sensitivity Experiments**

Several numerical experiments elucidated the sensitivity of the simulated isotopic field to temporal variations in the forcings and the modeling design. Table 2.2 outlines each experiment. Experiment 1 through 4 test the impacts of the seasonality of each forcing variables, $E$, $W$, $\bar{Q}$, and $P$, on the isotope variability. In the respective experiments, each forcing is replaced with
Figure 2.10: Distributions of the averaged contribution to isotopic variability: Daily $\Delta \delta w$ is divided into three components: (a) moisture flux-derived $(\Delta \delta w)_F$, (b) evaporation-derived $(\Delta \delta w)_E$, and (c) precipitation-derived $(\Delta \delta w)_P$. Absolute values of each delta are averaged, and the percentage of each derivation relative to the sum of three averaged absolute values is shown. Contour interval is 10%.
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<table>
<thead>
<tr>
<th>Experiment</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Averaged evaporation amount $\bar{E}$ for each grid</td>
</tr>
<tr>
<td>2</td>
<td>Averaged precipitable water $\bar{W}$ for each grid</td>
</tr>
<tr>
<td>3</td>
<td>Averaged moisture fluxes $\bar{Q} = (\bar{Q}<em>\phi, \bar{Q}</em>\psi)$ for each grid</td>
</tr>
<tr>
<td>4</td>
<td>Averaged precipitation amount $\bar{P}$ for each grid</td>
</tr>
<tr>
<td>5</td>
<td>Averaged daily forcings calculated from four 6-hourly forcings</td>
</tr>
<tr>
<td>6</td>
<td>$2.5^\circ \times 2.5^\circ$ rough spatial resolution design</td>
</tr>
</tbody>
</table>

an averaged value, $\bar{E}$, $\bar{W}$, $\bar{Q}$, and $\bar{P}$, for the simulation period in each grid. Global distribution maps of correlation coefficients between the time series fluctuations of the experiments and the control simulation are created to evaluated sensitivities. The results in Figure 2.11(a)-(d) show that the isotopic field is more sensitive to moisture flux $\bar{Q}$ and precipitation $P$ than to evaporation $E$ and precipitable water $W$. This indicates that temporal constant values of precipitable water and evaporation are acceptable to use for this simulation, particularly at low latitudes and in non-maritime regions.

Experiment 5 tests the sensitivity to the diurnal cycle of the forcings. All 6-hourly forcings from the GAME reanalysis are averaged to daily values. No significant difference is detected in the results shown in Figure 2.11(e). The use of the daily forcings instead of 6-hourly forcings is therefore acceptable.

Finally, in Experiment 6, increasing the model horizontal resolution to $2.5^\circ \times 2.5^\circ$ tests the sensitivity of results to horizontal resolution. Results are similar over the globe, as shown in Figure 2.11(f). Thus, $2.5^\circ \times 2.5^\circ$ spatial resolution is fine enough for simulation of isotopic circulation.

2.5.2 Precipitation from GPCP

Daily grid precipitation data from the GPCP (Global Precipitation Climatology Project) [Huffman et al., 2001] are used instead of the data from the GAME reanalysis. The data, originally $1^\circ \times 1^\circ$ are resampled to $1.25^\circ \times 1.25^\circ$. The suitability of using daily data instead of 6-hourly has been confirmed by the sensitivity experiments. Table 2.3 shows that both the correlation coefficients and the RMSEs relative to the observations at all three sites are improved. Figure 2.12 shows results for Chiangmai that compare with the control simulation and the observation. Isotopic depleted peaks on 1 July, 10 July, 10 August, 23 August, and 8 September and enriched peaks on 6 July, 12 July, and 13 August are better reproduced than the control simulation. The correlation coefficient has increased to 0.80 (from 0.76) with a significance level exceeding 99%, and the RMSE has decreased to 2.9%/o (from 4.2%/o). The better results yielded by the more accurate precipitation in GPCP indicate the more closely mass balance results the closer isotope variability towards reality, and securely support the remark in the last section, i.e., moisture transport at large scales controls the short-term isotopic variability of precipitation over Thailand.
Figure 2.11: Results of sensitivity experiments: Distribution of the correlation coefficients between the fluctuations in the experiments and the control simulation. (a) evaporation, (b) precipitable water, (c) moisture fluxes, and (d) precipitation, are respectively averaged for the simulated period in each sensitivity experiment. (e) all 6-hourly forcings are averaged to produce daily forcings. (f) a smaller spatial resolution \((2.5^\circ \times 2.5^\circ)\) is used. Contour interval is 0.1 for figure (b), 0.05 for figure (f), and 0.25 for the others.
Table 2.3: Results of the simulation using precipitation values from GPCP

<table>
<thead>
<tr>
<th>Location</th>
<th>Cor. Coef.</th>
<th>RMSE (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chiangmai</td>
<td>0.76 → 0.80</td>
<td>4.2 → 2.9</td>
</tr>
<tr>
<td>Sukhothai</td>
<td>0.74 → 0.77</td>
<td>4.1 → 2.8</td>
</tr>
<tr>
<td>Bangkok</td>
<td>0.56 → 0.60</td>
<td>3.5 → 2.8</td>
</tr>
<tr>
<td>Globe</td>
<td>0.76 → 0.75</td>
<td>4.8 → 5.2</td>
</tr>
</tbody>
</table>

Figure 2.12: Simulation results (black solid line) for Chianmai, using precipitation from GPCP, compared with the control simulation (gray-dashed line) and observations (bars).
Table 2.4: Results of the simulation using NCEP/NCAR data. (Note: precipitation is from GPCP and evaporation is from GAME reanalysis.)

<table>
<thead>
<tr>
<th></th>
<th>Cor. Coef.</th>
<th></th>
<th>RMSE (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Ave.</td>
<td>1998</td>
<td>1999</td>
</tr>
<tr>
<td>Chiangmai</td>
<td>0.62</td>
<td>0.69</td>
<td>0.55</td>
</tr>
<tr>
<td>Sukhothai</td>
<td>0.73</td>
<td>0.77</td>
<td>0.65</td>
</tr>
<tr>
<td>Bangkok</td>
<td>0.64</td>
<td>0.48</td>
<td>0.65</td>
</tr>
</tbody>
</table>

2.5.3 NCEP/NCAR Reanalysis

The GAME reanalysis includes only several months in 1998, and the simulated isotope circulation in this study was verified with observations made in 1998 only. To produce a longer model run, two years of the NCEP/NCAR reanalysis [Kalnay et al., 1996] are used. These daily data are provided on a global 2.5° × 2.5° grid. Only precipitable water $W$ and vertically integrated moisture fluxes $\tilde{Q}$ are calculated from the NCEP/NCAR data. Evaporation and precipitation are not included in the NCEP/NCAR data, so the averaged evaporation is taken from the GAME reanalysis and the daily precipitation is from GPCP. The suitability of these temporal and spatial resolutions and of using the averaged evaporation has been confirmed by the sensitivity experiments above.

Figure 2.13 shows the results of the simulation for Chiangmai, Sukhothai, and Bangkok for 1998–1999. Table 2.4 shows that the 2-year averaged correlation coefficients and RMSEs with observations remain at levels comparable to the control simulation. These figures and table suggest that the model accurately simulated interannual isotopic variability. For instance, the observations for July in 1999 show more enriched values than in 1998 (see Figure 2; most observed values are in the range of −4 and −12 in July 1998, 0 and −8 in July 1999), and the model reproduced the difference (see Figure 13; most predicted values are in the range of −8 and −16 in July 1998, −4 and −12 in July 1999). Obviously, there still exists the discrepancy of the consistent underestimation by the model, so that the model requires further improvement as suggested in the last section.

Figure 2.14 shows the global seasonal variability in $\delta^{18}O$ (weighted monthly mean for two years) compared to GNIP observations at eight sites that represent high latitude, temperate, subtropical, and tropical zones in the Northern and Southern Hemispheres. Correlation coefficients range from 0.50 to 0.94, which indicates good agreement with observations of regional isotopic characteristics, despite the large spatial variability. For all of the 389 GNIP sites, the correlation was a respectable 0.62. Results are not included for polar regions, which have been the main focus of many isotope-AGCM studies. Recall that the CFL condition for the model fails near the Poles. There is an obvious need for model improvement at very high latitudes.
Figure 2.13: The 2-year simulation results using NCEP/NCAR reanalysis compared with observations in 1998 (a, c, e) and 1999 (b, d, f) at Chiangmai (a and b), Sukhothai (c and d), and Bangkok (e and f).
Figure 2.14: Weighted monthly averages of the 2-year simulation using NCEP/NCAR (black diamonds) compared with the monthly observations of GNP (gray crosses) and the 2-year averaged monthly precipitation from GPCP (bars) at eight sites.
2.6 Summary and Conclusion

A new water isotope circulation model that includes a Rayleigh equation and the use of external meteorological forcings has been developed. A simulation for 1998 was forced with data from the GAME reanalysis. The results from this simulation were validated by GNIP observations of $\delta^{18}O$. Comparisons of monthly precipitation yielded a correlation coefficient of 0.76 with a significance level of >99%. The model also reproduced daily isotopic variability at Chiangmai, Thailand with $R=0.76$ and a significance level >99%. There was relatively poorer correlation with the rainfall amount at the site, so that 'amount effect' is insufficient to explain the short-term isotopic variability in precipitation. What controls the variability in the present model is the moisture transport system at large scales, in other words, reasonable water and isotopic mass balance on more than 100 km horizontal scale. The good reproduction of the model supports the dominance of the effect of the large scale moisture transport system on the short-term isotopic variability in Thailand during the rainy season. Regional processes of rain formation account for less of the isotopic variability.

A quantitative analysis of the short-term variability was made. The change in $\delta^{18}O$ in atmospheric vapor is caused by three factors: moisture flux, evaporation, and precipitation. At Chiangmai from April to October, 1998, moisture flux and precipitation both caused 37% of the isotopic variability; evaporation accounted for 26% of the variability. Globally, the contributions from each of the three factors have clear geographical distributions: moisture flux is most important in arid regions, precipitation dominates in regions of persistent rain, and evaporation is important in warm maritime regions. Globally, the averaged contribution of moisture flux was the largest, 46%. Thus moisture flux, in other words the transport and mixing of moist airmasses with different isotopic concentrations, is the prime generator of isotopic variability, and should be considered in similar studies.

Furthermore, sensitivity experiments revealed that the seasonality of precipitable water and evaporation had less impact on isotopic variability than did moisture flux and precipitation. The diurnal cycle and horizontal model resolution had little effect on the model results of isotopic variability. Therefore, at least daily moisture flux and precipitation and monthly or annual evaporation and precipitable water at 2.5° x 2.5° in the horizontal are required for acceptable simulation results of the Rayleigh-type isotope circulation model.

Then, the model was applied to two additional datasets. The more accurate GPCP precipitation dataset yielded improved model results at all three observation sites in Thailand. For example, at Chiangmai, the correlation increased to 0.80 from 0.76, and the RMSE decreased to 2.9‰ from 4.2‰. A 2-year simulation using the NCEP/NCAR reanalysis dataset produced very good agreements with the observed interannual variability in Thailand, and the simulated 2-year-averaged seasonality corresponded to GNIP observations at about 100 sites spanning the globe, with a correlation of 0.62.

Finally, we shall suggest some issues to be further progressed to diminish the discrepancy of the consistent over-depletion by the model. First of all, some lacking parts of isotopic physics in this model might have caused the underestimation. For example, evaporation from a falling droplet enriches the isotopic composition of surface precipitation to some extent. The second uncertainty is the use of the constant three types of evaporating isotopic ratio ($\delta e$). As Bosilovich and
Schubert [2002] show, a rate of land-evaporated water in precipitation (a water recycling ratio) is remarkably high in inland regions. $\delta e$ is obviously affected by precipitation isotopes while ’recycling’ in a regions. Even though the isotopic interactions between land and atmosphere have not yet reliably established as noted above, an isotopic evapotranspiration model by Gat and Matsui [1991] can possibly be a good model to be incorporated. Further, not only on land surface but also on sea surface, the spatial and temporal variability of $\delta e$ derived from the isotopic variability in sea surface water [Schmidt et al.,1999] shall be incorporated. Thirdly, the spatially uniform equilibrium fractionation scheme in the model includes uncertainty, too. In addition to the equilibrium fractionation, the kinetic fractionation that depends on the ambient temperature and humidity shall be taken into account. When we incorporate the fractionations to the model, the external forcings are used. Some simplification is required because of the limited number of variables, but the temperature and humidity at the representative condensation level can be obtained and they result appropriate kinetic and equilibrium fractionation by using the equations in Merlivat and Jouzel [1979] and Jouzel and Merlivat [1984]. It is required to consider the kinetic effects for both D and $^{18}$O to simulate variability of d-excess ($d = \delta D - 8 \times \delta^{18}O$).
Chapter 3

Evaluation of Two-dimensional Atmospheric Water Circulation Fields in Reanalyses by using Precipitation Isotopes Databases

Abstract:
A global Rayleigh-type H$_2^{18}$O circulation model was forced by two reanalysis products, the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis (NRA1) and the European Centre for Medium-Range Weather Forecasts (ECMWF) 15-year reanalysis (ERA15) datasets, and the global Network of Isotopes in Precipitation (GNIP) database was used to validate for 1979-1993 on monthly to inter-annual scales. We find out that the model monthly reproducibility is better with ERA15, particularly in Europe; the number of sites with a statistically $>99\%$ significant correlation is 87 out of 116 available European observational sites, whereas 4 sites are qualified with NRA1. Meanwhile, both monthly results are relatively similar and in good agreement with the observations in the rest of regions. Regardless of small number of comparable sites, we examine that inter-annual $\delta^{18}$O variability for 15 years is also better reproduced with the ERA15 in Europe, while there is quite similar reproducibility in the rest of regions, too. This study thus reveals that inconsistency of temporal variations in precipitation between the two re-analyses mainly, but indirectly, causes this difference of isotopic reproducibility. The isotope analyses diagnose accuracy of two-dimensional water circulation fields in datasets with a particular focus on precipitation processes, but in a slightly different manner over middle and high-latitude and over low-latitude: severer over former regions.

3.1 Introduction

Stable water isotopes (HDO and H$_2^{18}$O) in precipitation are tracers for the atmospheric hydrologic cycle, because their compositions are physically influenced by complex atmospheric
behavior (e.g., water vapor advection, condensation, or evaporation). Many observational studies have examined precipitation isotopes since Dansgaard [1964] [e.g., Welker, 2000; Stern and Blisniuk, 2002; Longinelli and Selmo, 2002; Kurita et al., 2003]. Interpretation is complicated, because simultaneous processes may cause similar isotopic variations [Burde and Zangvil, 2001].

Studies incorporating water isotopes into atmospheric general circulation models (AGCMs) have influenced general and global interpretations of precipitation isotopes [e.g., Jouzel et al., 1987; Hoffmann et al., 1998; Vugt et al., 2003]. Study results, however, are limited by the accuracy of AGCMs. Even though they have reasonably captured the major isotopic climatology, they seem to have difficulty reproducing “actual” short-term (event-based or daily) events [Hoffmann et al., 2000] or interannual isotopic signals and variability [Werner and Heimann, 2002].

Yoshimura et al. [2003] developed a simpler global model, namely a Rayleigh-type isotope circulation model, that uses a Rayleigh equation and external meteorological forcings (see Figure 2.3). The model reproduces “actual” daily H$_2^{18}$O variability over the sub-tropics, particularly Thailand, and monthly averages at global scales. Yoshimura et al. [2003] argued that moisture transport system at large scales is the main control of isotopic variability of precipitation. Thus, a reanalysis atmospheric dataset may yield a better estimate of global moisture transport than the self-generated atmospheric fields in AGCMs. They also highlighted the utility of isotopic analysis with the model: in case wrong hydrologic transport is simulated for the wrong reasons, isotopes can help in the diagnosis of problems.

Later, on the other hand, Lawrence et al. [2004] intensively surveyed on vapor and rain isotopes in the tropics, and noted better understanding of the cause of isotopic variability by the use of satellite and radar data combined with trajectory analyses. They indicated that precipitation or vapor flux convergence at and upwind from the sampling locations is the critical factor.

Above two arguments are substantially very similar even though their methods are different. The former uses the Eulerian transport method and quantitatively and implicitly computes the isotopic effect of upwind precipitation over the whole globe in one computation, while the latter adopted the Lagrangian approach to analyze the effect more directly but for some specific locations.

The control experiment in Yoshimura et al. [2003], however, was only seven-months long, which is not long enough to compare thoroughly with available observations. This study extends isotopic simulations by using two different reanalysis datasets to examine isotopic variability reproduction, particularly at inter-annual scales. Furthermore, reanalysis datasets were also compared to observed datasets.

The following section describes the isotope circulation model and data, including reanalyses and observations. The third section compares simulation results to observations. The fourth section identifies the characteristics of the reanalysis datasets that most impacted the simulated isotope fields. The last section provides a summary and conclusions.
3.2 Model Description and Simulation Design, and Observational Database

3.2.1 Model Description and Simulation Design

This study used a 2.5°×2.5° version of the isotope circulation model in Yoshimura et al. [2003]. The model is global (except for poleward regions of 85N and 85S; because of CFL condition), with one vertical layer. The model’s distinctive feature is its simplicity: Rayleigh isotopic fractionation employed with total vapor column and precipitation amount; constant isotopic values of evaporative flux; and complete mixing of vapor in a grid-box every model time step. In addition to, and in contrast to these simplifications, the model uses forcing variables from external meteorological datasets to reproduce realistic water cycle.

Global water and H$_2^{18}$O cycles are computed by maintaining the atmospheric water balance (3.1) [Oki et al., 1995] and the isotopic mass balance (3.2) [Yoshimura et al., 2003], respectively, with an upstream scheme at each grid point at every time step

$$\frac{\partial W}{\partial t} = - \nabla \cdot \vec{Q} + (E - P)$$  \hspace{1cm} (3.1)

$$\frac{\partial (\delta wW)}{\partial t} = - \nabla \cdot (\delta w\vec{Q}) + \delta eE - \delta pP$$  \hspace{1cm} (3.2)

where $W$, $\vec{Q}$, $P$, and $E$ are the forcing variables, and represent precipitable water, vertically integrated horizontal vapor flux vector (consists of zonal and meridional fluxes), precipitation and evaporation, computed from reanalyses, respectively; and where $\delta w$, $\delta p$, and $\delta e$ are the isotope ratios of precipitable water, precipitation, and evaporation, respectively.

For the forcing variables, $W$, $\vec{Q}$, $P$, and $E$, two sets were calculated from the following two reanalysis datasets: the National Centers of Environmental Prediction and National Center for Atmospheric Research (NCEP/NCAR) reanalysis (NRA1) dataset [Kalnay et al., 1996] and the European Centre for Medium-Range Weather Forecasts (ECMWF) 15-year reanalysis (ERA15) dataset [Gibson et al., 1997]; the resolutions of both reanalyses were converted to 2.5°×2.5° and 6-hourly for 0000 UTC 1 January 1979 to 1800 UTC 31 December 1993. Global isobaric wind speeds and specific humidity fields were used to calculate precipitable water, $W$, and vertically-integrated vapor flux, $\vec{Q}$. Precipitation is given as the sum of convective precipitation and large-scale condensation.

For isotopic value of evaporative flux, $\delta e$, the model uses three constant values in time series as same as Yoshimura et al. [2003]: −9.4‰ on seas, −10‰ on land between 40N and 40S, and −15‰ on other land because isotopic variability caused by the moisture transport system is mainly focused in this study rather than the effect of isotopic variability of evaporative source. This simplification perhaps leads biases to some extent, particularly systematic underestimation (see below).

For the Poles, the model does not calculates but assign a constant value for polar vapor isotopes ($\delta w_p = −30$ ‰) in poleward of 85N or 85S. This might lead other biases as described below, too.
3. REANALYSES EVALUATION BY ISOTOPES

<table>
<thead>
<tr>
<th>Table 3.1: Specifications of the two simulations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Value/Description</td>
</tr>
<tr>
<td>-------------</td>
</tr>
<tr>
<td><strong>General</strong></td>
</tr>
<tr>
<td>Spatial resolution</td>
</tr>
<tr>
<td>Simulation period</td>
</tr>
<tr>
<td>Model time step</td>
</tr>
<tr>
<td><strong>Meteorological Forcing Variables</strong></td>
</tr>
<tr>
<td>$W$ (precipitable water)</td>
</tr>
<tr>
<td>$\tilde{Q}$ (vapor flux)</td>
</tr>
<tr>
<td>$E$ (evaporation)</td>
</tr>
<tr>
<td>$P$ (precipitation)</td>
</tr>
<tr>
<td><strong>Isotopic Parameter</strong></td>
</tr>
<tr>
<td>$\alpha$ (fractionation factor)</td>
</tr>
<tr>
<td>$\delta E$ (isotope ratio of $E$)</td>
</tr>
<tr>
<td></td>
</tr>
<tr>
<td>$\delta w_p$ (polar vapor isotope)</td>
</tr>
</tbody>
</table>

Then, the model computes $^{18}$O in precipitable water and precipitation by using Rayleigh equation [Yoshimura et al., 2003]

$$
\delta w = \left( \left( \frac{W}{W^*} \right)^{\alpha-1} (1 + 10^{-3} \delta w^*) - 1 \right) \times 10^3
$$

$$
\delta p = \frac{\delta w^* - \delta w W}{\left( 1 - \frac{W}{W^*} \right)}
$$

(3.3)

where $W$ and $\delta w$ are the precipitable water and its isotopic ratio, respectively; $W^*$ and $\delta w^*$ are those just before precipitation occurs ($W^* = W + P \Delta t$); $\delta p$ is isotopic ratio of precipitation; and $\alpha$ is the fractionation factor.

The two sets of forcing variables were used to drive the model, namely “NRA1 run” and “ERA15 run.” The model time step, $\Delta t$, is 10 minutes; the $^{18}$O in precipitation is weighted by the precipitation amount to yield a daily averaged value. Monthly weighted averages or annual averages can be computed from daily values. Specifications for the two simulations are displayed in Table 3.1.

3.2.2 Observational Database

The Global Network of Isotopes in Precipitation (GNIP) database (International Atomic Energy Agency (IAEA) and World Meteorological Organization (WMO), 2001, accessible online at http://isohis.iaea.org) was used to compare simulated $\delta^{18}$O in precipitation with observed $\delta^{18}$O. The observational data are reported as monthly values. Model results were carefully compared to corresponding observations.
3.3 Results

3.3.1 Monthly Results and Validation

Comparisons in Vienna and Bangkok

Figure 3.1 and 3.2 show monthly precipitation $\delta^{18}O$ variations at Vienna and Bangkok for 1979-1993 from NRA1 run and ERA15 run, comparing with GNIP monthly observations. Vienna and Bangkok were used to represent Europe and southern Asia; at both sites observational data are available for all 15-year period. Moreover, Yoshimura et al. [2003] validated daily data at Bangkok for 1998 and 1999.

At Vienna (Figures 3.1a and 3.2a), observed precipitation $\delta^{18}O$ decreased in winter and increased in summer. The amplitude of seasonal variation varied from year to year, but was about 10 $\%_o$ on average. In the NRA1 run, however, simulated precipitation $\delta^{18}O$ was steady around -10 $\%_o$ and the seasonal cycle was not reproduced. In contrast, the ERA15 run successfully reproduced seasonality, although the seasonal cycle amplitude was smaller, owing to underestimation of summer precipitation $\delta^{18}O$. The two runs showed quite different simulated monthly precipitation $\delta^{18}O$ at Vienna; isotopic seasonality was more realistic in the ERA15 run. The correlation coefficient ($R$) of the ERA15 run was 0.69 and thus statistically significant at a level exceeding 99%. In contrast, the null hypothesis cannot be rejected with $R = 0.03$ for the NRA1 run.

At Bangkok, NRA1 and ERA15 showed similar results. Both reflected strong seasonality, i.e., depletion in summer and enrichment in winter. This seasonality was echoed in the observations, which correlated to NRA1 and ERA15 at significance levels exceeding 99% (0.69 and 0.70, respectively). However, simulated values were generally smaller. The mean bias for comparable data was $-5.6 \%_o$ and $-5.1 \%_o$ in the NRA1 and ERA15 runs, respectively.

This systematic bias results from simplifications in the model, most likely due to underestimation of isotopic values of evaporative flux from seas ($\delta e = -10 \%_o$). For instance, Craig and Gordon [1965] assumed a closed isotopic balance between evaporation the mean precipitation ($\delta e = -4 \%_o$). Afterwards, isotope-AGCM studies, particularly Jouzel and Koster [1996] argued limitation of the closure assumption and recommended the use of isotope-AGCM output. However, isotopic composition of marine evaporation should have variability in a range more than $-10 \%_o$ according to ECHAM-iso [Hoffmann et al., 1998] output.

Again, this study aims to capture isotopic variability caused by the moisture transport system. Therefore, it adopts constant isotopic values for evaporative flux; nevertheless the model reproduced the variability of precipitation isotopes. Hence, it is arguably noted that moisture transport system, including fractionation derived by upwind precipitation, horizontal mixing by vapor flux, and supply by evaporation, controls isotopic fields of vapor and precipitation.

Global Validation of Monthly Results

The global maps in Figure 3.3 show the correlation coefficient distributions for all available observation sites. The GNIP observational data have various periods so that the number of data during the simulation differs for each site as shown in Table 3.2. Furthermore, even if
Figure 3.1: Fifteen-year variations in monthly precipitation $\delta^{18}O$ from NRA1 (gray crosses with gray lines), ERA15 (open circles with dashed lines), and available G NIP observations (black diamonds with solid lines). (a) Vienna and (b) Bangkok
Figure 3.2: Scatterplots of Figure 3.1; monthly precipitation $\delta^{18}O$ from NRA1 (crosses) and ERA15 (circles) runs are compared with corresponding GNIP observed data. The gray line and black dashed line indicate linear regression of the NRA1 and ERA15 runs, respectively. (a) Vienna and (b) Bangkok.
Table 3.2: Available monthly data in GNIP observations for the 15-year period categorized by regions. The seven regional abbreviations are given in Figure 3.

<table>
<thead>
<tr>
<th></th>
<th>&gt; 80%</th>
<th>50~80 %</th>
<th>20~50 %</th>
<th>5~20 %</th>
<th>Total</th>
<th>(others)</th>
</tr>
</thead>
<tbody>
<tr>
<td>WE</td>
<td>31</td>
<td>17</td>
<td>31</td>
<td>37</td>
<td>116</td>
<td>(172)</td>
</tr>
<tr>
<td>EE</td>
<td>1</td>
<td>4</td>
<td>24</td>
<td>27</td>
<td>56</td>
<td>(54)</td>
</tr>
<tr>
<td>AF</td>
<td>0</td>
<td>0</td>
<td>5</td>
<td>6</td>
<td>11</td>
<td>(26)</td>
</tr>
<tr>
<td>OC</td>
<td>2</td>
<td>5</td>
<td>1</td>
<td>3</td>
<td>11</td>
<td>(7)</td>
</tr>
<tr>
<td>NA</td>
<td>1</td>
<td>0</td>
<td>13</td>
<td>7</td>
<td>21</td>
<td>(32)</td>
</tr>
<tr>
<td>SA</td>
<td>1</td>
<td>4</td>
<td>17</td>
<td>34</td>
<td>56</td>
<td>(37)</td>
</tr>
<tr>
<td>MR</td>
<td>2</td>
<td>5</td>
<td>6</td>
<td>7</td>
<td>20</td>
<td>(40)</td>
</tr>
<tr>
<td>Total</td>
<td>38</td>
<td>35</td>
<td>97</td>
<td>121</td>
<td>291</td>
<td>(368)</td>
</tr>
</tbody>
</table>

observational data exist, simulated monthly precipitation $\delta^{18}$O may be missing if the monthly precipitation in the reanalysis is zero. Figure 3.3 shows all sites with more than nine pairs of comparable observed and simulated data (5% of all 180 months). Both the NRA1 (Figure 3.3a) and ERA15 runs (Figure 3.3b) have 291 such sites.

A clear difference is evident over western Eurasia (WE; see Figures 3.3). In contrast, there is little difference over other regions, except in mid-Eurasia (the northwestern part of region EE: eastern Eurasia). As shown in Figures 3.4, 87 out of 116 sites correlate with observations at the 99% significance level for the ERA15 run in WE; only 4 out of 116 show such correlation levels in the NRA1 run. There is little difference between the two runs in EE and elsewhere, where results from both models agree well with observations.

In contrast, Figure 3.5 shows global distributions of systematic error (mean bias). In a global sense, there is little difference in mean bias between two results, except for WE region, where random errors offset systematic biases. In most sites in both runs, some extent of underestimations are notably seen. Underestimation of isotopic ratio of marine evaporative sources, as described above, is one possible cause. Larger biases in regions near seas (central America and western Europe) than in inland (Brazil and eastern Europe) can perhaps support this idea. However, this cannot explain all results, for example larger negative biases in the inland of south China. This region, “continental cycling ratio” (land-originated precipitation) is the highest level [Yoshimura et al., 2004; Bosilovich et al., 2002]. Thus too low value for land-originated evaporative source may produce this problem. In these regards, reasonable estimates of evaporative isotopes would be expected for further improvement of the model. Furthermore, in northern region of NA, there are positive biases; only in this region, particularly west of Hudson Bay, northerly vapor transports are dominant in both reanalyses, resulting inflow of too enriched polar vapor, i.e. $\delta w_p = -30 \%_0$.

3.3.2 Annual Results and Validation

In this section, the reproducibility of inter-annual variability is examined using weighted annual averages from which seasonality has been eliminated. The weighted annual $\delta^{18}$O was
Figure 3.3: Global distribution of correlation coefficients between GNIP observations and simulated predictions using monthly data: results from the 15-year run with (a) NRAI and (b) ERA15. Circles indicate observational sites with statistically significant correlations at a level exceeding 99%, triangles denote a level exceeding 90%, and crosses show sites where the null hypothesis cannot be rejected at the 90% confidence level. Regions within 1000 km of any site are shaded. Regional abbreviations represent North America (NR), South America (SA), Western Eurasia (WE), Eastern Eurasia (EE), Africa (AF), Oceania (OC), and Maritime regions (MR).
3. REANALYSES EVALUATION BY ISOTOPES

Figure 3.4: Number of sites in Western Eurasia (WE), Eastern Eurasia (EE), and other regions categorized by correlation significance levels (more than 99%, 95%, 90%, and the other) between the GNIP observations and (a) the NRA1 or (b) ERA15 runs.

Calculated from GNIP monthly observations only where observed monthly totals exceeded 10 mm. Precipitation totals of less than 10 mm were assumed to be zero in the calculation. This procedure avoided gaps in the observations but did not affect annual results, because weighted annual $\delta^{18}$O strongly depends on the precipitation in wet months. However, many gaps in the GNIP observations remained. Only 56 sites had more than five annual precipitation $\delta^{18}$O data in this 15-year period. Of these, 38 sites were located in WE.

Figure 3.6 shows time series of 15-year interannual variations in weighted $\delta^{18}$O annual anomalies from the GNIP observations, NRA1 run, and ERA15 run at eight sites, which represent each region shown in Figure 3.3. The eight sites are Bad Salzuflen, Germany, and Vienna, Austria (WE region); Ottawa, Canada (NA region); Midway Island, USA (MR region); Hong Kong, China, and Bangkok, Thailand (EE region); Brisbane, Australia (OC region); and Buenos Aires, Argentina (SA region). No site is available for region AF. Interannual variations of precipitation amounts from the three data sources are also shown.

At Bad Salzuflen and Vienna, interannual variations from the ERA15 run agreed with observations, with correlations of 0.95 and 0.81 (statistically significant at a level >99%), respectively. Variations in the NRA1 run did not agree with observations. At Ottawa, neither result resembled the observations. Even though the ERA15 run ($R = 0.11$) had a lower correlation than the NRA1 run ($R = 0.50$) at Ottawa, the ERA15 run better simulated the enrichment peak in 1987. At Midway, both runs traced observed interannual variability with similar correlations ($R = 0.69$ and 0.83 for the NRA1 and ERA15 runs, respectively). At Hong Kong and Bangkok, results from the two models were also similar, although the correlation coefficients from the NRA1 run were slightly higher at both sites. At Brisbane, results for both models were similar.
Figure 3.5: As in Figure 3.3, but for mean bias. Solid and dashed circles indicate negative biases (underestimation) and positive biases (overestimation), respectively.
to observations. In Buenos Aires, the ERA15 run reproduced interannual variability better than the NRA1 run. However, only five years of comparable data exist for Buenos Aires.

Figure 3.6 also shows better interannual variations of precipitation in ERA15 at two European sites as same as Stendel and Arpe [1997] found. In Figure 3.7, precipitation fields from the reanalyses are compared to an observation-based pentad precipitation by the Global Precipitation Climatology Project (GPCP) [Huffman et al., 1997]. Precipitation from ERA15 is more realistic than precipitation from NRA1 over large areas. Better precipitation may lead (or be derived from) more realistic hydrologic cycle, and should result better isotopic simulation. Indeed, better simulated precipitation isotopes in ERA15 run correspond over western Eurasia (including Europe and Russia), east of North America, and the Caribbean. However, other regions, including north edge of North America and northeast of South America, ERA15 run did not show clear differences. It may be possible to note that inaccurate precipitation over northeast of South America in both reanalyses result no improvement in either of the isotopic simulations. In these regards, we further discuss on what made the two different isotopic simulations in the next section.

3.4 Discussion — What Made Different Isotopic Simulations?

Figure 3.8 compares daily variations in atmospheric fields and simulated precipitation $\delta^{18}$O by the NRA1 and ERA15 runs for 1979 at Vienna and Bangkok. The figure highlights the cause of similarities and differences between the two simulated isotopic fields. Atmospheric fields include precipitable water, $W$; vapor fluxes, $Q_x$, (zonal) and $Q_\phi$, (meridional); vapor convergence $- \nabla \cdot \bar{\dot{Q}}$, precipitation, $P_i$ and evaporation, $E$. Even though analytic fields, i.e., precipitable water and vapor fluxes (zonal and meridional), are very similar in both reanalyses, deduced convergence have some differences and presumably forecasted products (i.e., precipitation and evaporation) become very different. The differences in precipitation and evaporation between the two reanalyses seemingly force the different isotopic results at the two sites.

Figure 3.9 displays 15-year averaged correlation of simulated daily precipitable water $\delta^{18}$O ($\delta w$) fields, (a) annual results, (b) northern summer (JJA), and (c) northern winter (DJF), using the two runs. In low-latitude (between 30°N and 30°S), particularly in monsoonal regions, correlations in the annual results become emphasized than JJA and DJF, because of clear seasonal changes. In contrast, in middle and high-latitude, correlations remain at the same level or even become lower in western Eurasia. Near the north pole, high correlations appear in JJA, but it is owing to the model limitation near the Poles, i.e., northerly vapor transport inflows the constant pole vapor isotopes ($\delta w_p$).

In low-latitude (30°N–30°S), results indicate similarity in simulated isotopic fields at continental subtropics (between 10° to 30° in both hemispheres), but the Equatorial regions and the maritime regions including ITCZ and SPCZ show low correlations. In particular, very low values appear over northeast of southern Pacific. These spatial correlation patterns correspond with those of precipitable water (also those of precipitation and evaporation) between two reanalyses (Figure 3.10), and interestingly, those between NRA1 and observation in Trenberth and Guillemot [1998]. It presumably reflects that the not-well-defined tropical structures in both
Figure 3.6: captioned on the next page
Figure 3.6: (figures on the previous page) Interannual anomaly variability of weighted annual averages using NRA1 (gray crosses with gray lines) and ERA15 (open circles with dashed lines) compared with the same inter-annual anomaly variability from the GNIP observations (black diamonds with solid lines). Annual precipitation amounts in NRA1 (light gray bars) and in ERA15 (medium gray bars) compared with GNIP ground-based observations of precipitation (dark gray bars) are also shown. Note that observed annual data are not always available for all 15 years; some months have missing observations. (a) Bad Salzuflen, Germany; (b) Vienna, Austria; (c) Ottawa, Canada; (d) Midway Island, USA; (e) Hong Kong, China; (f) Bangkok, Thailand; (g) Brisbane, Australia; and Buenos Aires, Argentina (h).

Figure 3.7: Comparisons of precipitation fields of NRA1 and ERA15 with GPCP. Correlations of pentad precipitation amount for 1979–1993 between the reanalyses and GPCP for (a) NRA1, (b) ERA15, and (c) difference of two correlations (ERA15 minus NRA1). Black shading indicates no data in the GPCP.
reanalyses resulted large variety of hydrologic cycle, particularly similar (and likely reasonable) cycle over continents and different (and robust) over oceans, leading to corresponded simulation isotopic fields. A lack of observational sites with plentiful number of precipitation isotope data makes discussion on whether ERA15 simulated better or not over these oceans difficult. In the future, the collection of dense and continuous observations is important.

In contrast, over middle and high-latitude regions (> 30°), correlation is generally at lower level than low-latitude: large differences appear over maritime regions, particularly over the southern ocean, and over continents including eastern Eurasia, Greenland, and the east and west coasts of North America. Over these regions, even though precipitable water and evaporation in both reanalyses are very similar (Figures 3.10), simulated isotopic fields did not become similar. This is owing to two possible reasons. First, precipitation (and precipitable water, vapor fluxes, and evaporation, even though their differences are relatively small) has more sensitivity on isotopic results over these regions than low-latitude. Second, vapor isotopes are already influenced by accumulated fractionation effect with precipitation upwind to the regions; precipitation over middle and high-latitude are possibly influenced by isotopic variations over low-latitude, resulting relatively different simulated isotopic fields.

It is thus valid to argue that isotope simulations and their analyses with observations can diagnose differences in the moisture transport system (in other words, hydrologic cycle) owing to differences in precipitation processes. It might be useful for diagnoses of global or regional hydrologic dataset (e.g. reanalyses, GCM output, observations, etc.) However, this diagnoses of isotopes should be carefully used. Reproducibility of precipitation $\delta^{18}O$ is evaluated in a slightly different manner over low-latitude and that over middle and high-latitude. Over middle and high-latitude, isotopic reproducibility tends to be lower, but more sensitive to precipitation fields, than that over low-latitude. In short, the isotopic evaluation is severer over middle and high-latitude. Discussion on underlying processes that make different precipitation in ERA15 and NRA1 should be interesting, but it would most likely lead us to issues on complicated and sophisticated parameterizations of their respective GCMs. Therefore, further inquiry for parameterization issues is out of the main scope and it would not be discussed in this study.

3.5 Summary and Conclusion

Two reanalyses, the NCEP/NCAR reanalysis (NRA1) and the ECMWF 15-year reanalysis (ERA15), were applied to an isotope circulation model [Yoshimura et al., 2003] to determine model accuracy and examine the quality of forced reanalyses. The global network of isotopes in precipitation (GIP) data validated 15-year simulations (1979–1993) at monthly and inter-annual time scales. The ERA15 dataset reproduced monthly isotopic variations better, particularly in Europe, where 94 out of 116 sites correlated with observations at levels exceeding 99%; only four sites had the same correlation levels in the NRA1 results. Over the rest of the globe, little difference was evident in simulated isotope fields; both models adequately reproduced variability of the observed precipitation $\delta^{18}O$.

There remained systematic biases to some extent. These seem owing to the model simplifications; too simple evaporative isotopic values caused underestimations and the model’s technical
Figure 3.8: Daily variations in atmospheric fields (a1-a6, b1-b6) and simulated precipitation $\delta^{18}O$ (a7, b7) in the NRA1 and ERA15 runs for 1979 at Vienna (a1-a7) and Bangkok (b1-b7). The atmospheric fields include precipitable water, $W$ (a1, b1); zonal vapor flux, $Q_\lambda$ (a2, b2); meridional vapor flux, $Q_\phi$ (a3, b3); vapor convergence, $-\nabla \cdot Q$ (a4, b4); precipitation, $P$ (a5, b5); and evaporation, $E$ (a6, b6).
Figure 3.9: Fifteen-year averaged correlation coefficients of daily precipitable water $\delta^{18}O$ between the NRA1 and ERA15 runs. For (a) all 15 years, (b) northern summer (JJA), and (c) northern winter (DJF).
Figure 3.10: Global distributions of 15-year averaged correlation coefficients between NRAI1 and ERA15 for daily (a) precipitable water, (b) precipitation, and (c) evaporation.
limitation near the Poles caused overestimation.

Interannual variability in precipitation isotopes was also examined, although only a relatively small number of stations could be compared to the models at longer timescales. Only 58 sites had more than 5 years of annual observed data in the 15 years. Similar to the monthly results, however, the ERA15 simulation matched observations better over Europe; over the rest of the globe, ERA15 and NRA1 yielded similar agreements with observations.

Spatiotemporal distribution of precipitation from ERA15 are more accurate than that from NRA1 over large areas, and it lead more reasonable hydrologic cycle, resulting the better ERA15 isotopic simulation over western Eurasia (including Europe and Russia), east of North America, and the Caribbean. However, other regions with better precipitation, including north edge of North America and northeast of South America, ERA15 run did not show clear isotopic differences. A possible reason is that too inaccurate precipitation cannot improve the isotopic simulations.

Isotope simulations and their analyses with observations reflect accuracy of the moisture transport system (i.e., hydrologic cycle) with a focus on correctness of precipitation processes. It might be useful for diagnoses of global or regional hydrologic dataset (e.g. reanalyses, GCM output, observations, etc.) However, reproducibility of precipitation δ18O should be carefully evaluated over low-latitude and middle and high-latitude: over middle and high-latitude, isotopic reproducibility tends to be lower, but more sensitive to precipitation fields, than that over low-latitude; in short, the isotopic evaluation is severer over middle and high-latitude.
Chapter 4

Iso-MATSIRO, a Land Surface Model that Incorporates Stable Water Isotopes

Abstract:
This study describes Iso-MATSIRO, a land surface model that includes stable water isotopes and simulates physically reasonable isotopic fluxes and reservoirs at the ground. The model calculates kinetic and equilibrium fractionation of HDO and H$_2$O between ice, liquid, and vapor phases and separately considers soil surface evaporation, vegetation transpiration, evaporation from the canopy-intercepted reservoir, and snow sublimation. One-dimensional simulations with modeled meteorological forcings showed plausible features in the annual isotopic budget, seasonal variations of $\delta^{18}O$ in soil moisture, diurnal variations of leaf water with some enrichment, and $\delta$-diagram of representative surface reservoirs and fluxes. A subsequent, independent global simulation used Iso-MATSIRO coupled with an atmospheric isotope circulation model for a half year in 1998. Simulated precipitation $\delta^{18}O$ was closer to observations than in a previous study, confirming the physical treatments of isotopes in the land surface processes.

4.1 Introduction

Stable isotopes of water (HDO and H$_2$O) are good tracers for the hydrologic cycle because their concentrations in water reflect the accumulated record of physical phase changes. In particular, heterogeneity in precipitation isotopes in time and space reflects complex atmospheric behavior. Many observational studies of precipitation isotopes have occurred since Dansgaard [1964], and a variety of models can explain observed precipitation isotopes [e.g., Gat and Matsui, 1991]. Most models, however, excessively simplify or even neglect isotopes of vapor evaporated from the land surface even though the isotopes in land evaporation will reflect precipitation and complex land surface processes. For example, the atmospheric general circulation models (AGCM) in Hoffmann et al. [1998] incorporated isotopes but assumed no isotopic fractionation from the land surface. Yoshimura et al. [2003] assumed constant isotopic compositions from
evaporative sources though noted that this method introduced discrepancies between observation and simulation results. Moreover, Bosilovich et al. [2000] and Yoshimura et al. [2004b] showed a quantitatively large contribution of land surface evaporation to land precipitation. Koster et al. [2004] extracted “hot spots” of land surface which have large influences to atmosphere, and Henderson-Sellers et al. [2002] pointed out that the climatic shift of precipitation isotopes over Amazonian forest could not be reasonably explained without the land processes.

Precise isotopic mechanisms in land surface processes have been studied mainly in a view of ecosystem–biological researches regarding carbon dioxide exchange with carbon isotope (13C) in CO2 (e.g., Yakir and Sternberg [2000], Farquhar and Gan [2003], etc.) Water isotopes are also regarded interesting and have been studied in regards of land evaporation processes (e.g., Brunel et al. [1992], Martinelli et al. [1996], Brunel et al. [1997]Wang et al. [1998], He and Smith [1999], Wang and Yakir [2000], etc.)

Recently, a few land surface models that incorporate stable water isotopes have been developed [Riley et al., 2002; Riley et al., 2003; McGuffie and Henderson-Sellers [2004], Braud et al., 2005a; Braud et al., 2005b]. Furthermore, the first intercomparison project of water isotope embedded land surface models (iPILPS: isotopes in the project for intercomparison of land-surface parameterization schemes) started in 2004 [Henderson-Sellers, 2005; Henderson-Sellers et al., 2005]. The present study develops a land surface model that incorporates isotopes and yields reasonable estimates of isotopes for evaporation from land sources under various atmospheric conditions.

This study used the MATSIRO (Minimal Advanced Treatments of Surface Interaction and RunOff; Takata et al. [2003]) land surface model (LSM). The LSM was modified for energy and water flux partitioning in PILPS-2(e) [Bowling et al., 2003] and optimized to treat water and energy fluxes in a large-scale global simulation. The model includes explicitly separated treatments of evapotranspiration that can show different isotopic fractionation in evaporation from soil, transpiration from vegetation, evaporation from intercepted water by canopy, and sublimation from snow, with snow-free or snow-covered distinctions. Vertical soil water transport and advective water runoff are explicitly modeled with physical equations.

The following section details model structure and equations. Results from one-dimensional simulations in the vertical at three locations are given in the third section. Following section describes an independent global simulation in which Iso-MATSIRO is coupled with an atmospheric isotope model. Simulated precipitation isotopes are compared to global observations and previous simulations. The last section includes a summary and conclusions.

### 4.2 Model description

This section describes the original land surface model, MATSIRO, the isotopic physics adopted in the model, and modules integrated with MATSIRO to simulate HDO and H218O of all water fluxes and reservoirs. Isotopic values are expressed as parts per thousand (‰) relative to standard mean ocean water (SMOW).
4.2.1 MATSIRO

MATSIRO [Takata et al., 2003] was developed for global and regional climate studies. MATSIRO has a single-layer canopy and albedo; bulk coefficients are evaluated based on a multilayer canopy model. Fluxes are calculated from the energy balance at the ground and canopy surfaces in both snow-free and snow-covered portions that consider the subgrid snow distribution. Evaporation from canopy interception and transpiration on the basis of photosynthesis are included. A simplified TOPMODEL [Beven and Kirkby, 1979] calculates baseflow runoff, in addition to surface flows. Snow has up to three layers depending on the snow water equivalent, and snow layer temperatures are calculated with thermal conduction equations. Snowmelt and refreeze are considered. There are five soil layers in which energy and water movements are treated with physical equations that consider freezing and condensation. Model results are described in Hirabayashi et al. [2005] and Miyazaki et al. [manuscript in preparation]. Sakamoto et al. [2004] describes a simulation coupled with an AGCM. These studies obtained satisfactory results.

In Iso-MATSIRO, all variables that contain water amount or flux have respective isotopic concentrations. Isotopic concentrations that satisfy isotopic mass balances are computed at each time step. Figure 4.1 schematically shows the model structure of Iso-MATSIRO.

![Figure 4.1: Schematic representation of Iso-MATSIRO](image-url)
4. DEVELOPMENT OF ISO-MATSIRO

4.2.2 Basic surface isotopic fluxes with fractionation

When phase changes (e.g., liquid to gas) at any surface (i.e., soil, stoma, canopy, and snow) occur in the model, kinetic isotopic fractionation of the isotopic composition in the surrounding vapor is considered. Following Craig and Gordon [1965], turbulent diffusion is assumed to produce no isotopic fractionation, although molecular diffusion does. Surface flux $F$ and corresponding isoflux $F^{iso}$ are expressed as

$$F = \rho C_F |V_s| \left( q^*(T) - q_a \right)$$

(4.1)

$$F^{iso} = \rho C_{F,iso} |V_s| \left( q^*(T) R_\alpha(T) - q_a R_a \right)$$

(4.2)

$$R_F \equiv \frac{F^{iso}}{F} = \frac{C_{F,iso}}{C_F} \left[ \frac{q^*(T) R_\alpha(T) - q_a R_a}{q^*(T) - q_a} \right]$$

(4.3)

where $R_F$ is the isotopic ratio of flux $F$; $\rho$ is concentration of water; and $C_F$ and $C_{F,iso}$ denote bulk coefficients of water and isotopes against water vapor, respectively, where surface roughness resistance is considered. The ratio of bulk coefficients, $C_{F,iso}/C_F$ is the kinetic fractionation coefficient, and the approximate values in Merliviat and Jouzel [1979], which are functions of wind speed, are used. $V_s$ indicates wind speed on the surface, and $q^*(T)$ and $q_a$ are the saturated specific humidity at temperature $T$ and the specific humidity of the surrounding vapor, respectively. $R$ and $R_a$ denote the isotopic composition of water in the surface reservoir and surrounding vapor, respectively. $T$ is temperature at the reservoir, and $\alpha$ is equilibrium fractionation factor obtained from Majoube [1971a].

$R_F$ is isotopic composition of water flux, and all model calculations that manipulate isotopic mass balance do so considering isoflux ($F^{iso} = R_F F$). Therefore, isotopic exchange is not considered when net water flux is zero.

4.2.3 Canopy

Iso-MATSIRO describes canopy evaporation flux, $E_c$, and its isotopic composition, $R_{E_c}$, as follows:

$$E_c = \begin{cases} \int_{can} \rho C_{H_c} |V_a| \left( q^*(T_c) - q_a \right) & (q^*(T_c) - q_a) > 0 \\ \rho C_{H_c} |V_a| \left( q^*(T_c) - q_a \right) & (q^*(T_c) - q_a) \leq 0 \end{cases}$$

(4.4)

$$R_{E_c} = \begin{cases} \frac{\alpha_k q^*(T_c) R_c - q_a R_a}{q^*(T_c) - q_a} & E_c > 0 \\ \frac{R_a / \alpha^*(T_a)}{E_c \leq 0} \end{cases}$$

(4.5)

where $\int_{can}$ is wet area rate of the canopy, $C_{H_c}$ is the bulk energy coefficient from canopy to atmosphere, $R_c$ is the isotopic ratio of canopy reservoir, and $\alpha^*$ denotes an equilibrium fractionation factor for liquid to vapor ($\alpha$) when $T_c$ exceeds 0°C and for ice to vapor ($\alpha_{ice}$) when $T_c$ is cooler than 0°C. Note that dew condenses onto the entire canopy regardless of canopy wetness. Iso-MATSIRO is “big-leaf” type model and canopy intercepted water is one reservoir. Canopy reservoir water $C$ and its isotopic ratio $R_c$ are calculated with the above evaporation flux and isoflux as

$$\frac{\partial C}{\partial t} = I_c - E_c - D_c$$

(4.6)
\[
\frac{\partial CR_c}{\partial t} = I_c R_t - E_c R_e - D_r R_c \tag{4.7}
\]

where \( I_c \) is input flux for the canopy reservoir. Flux is defined using total precipitation flux and leaf area index (LAI); \( D_r \) is the dripping flux from the canopy, and \( R_I \) is the isotopic ratio of \( I_c \).

### 4.2.4 Snow

Sublimation from snow, \( E_{sn} \), and its isotopic composition, \( R_{E_{sn}} \), are defined as

\[
E_{sn} = A_{sn} \rho C_{H_s} |V_o| (q^\circ(T_s) - q_a) \tag{4.8}
\]

\[
R_{E_{sn}} = \begin{cases} 
\frac{q^\circ(T_s) R_m - q_a R_s}{\alpha_{icr}(T_s)} & \text{if } E_{sn} > 0 \\
\frac{R_a}{\alpha_{icr}(T_a)} & \text{if } E_{sn} \leq 0
\end{cases} \tag{4.9}
\]

where \( A_{sn} \) is the subgrid snow fraction, \( C_{H_s} \) is the bulk energy coefficient from ground surface to atmosphere, \( \alpha_{icr} \) is fractionation factor between ice and vapor, and \( R_{sn} \) is the isotopic ratio of all snow.

Iso-MATSIRO allows up to three snow layers, tracks the amount of snow water equivalent (SWE), and considers rain freezing, snow melting, and refreezing of melted snow in addition to snowfall. The model assumes an equilibrium fractionation for these phase changes. Equilibrium-fractionated meltwater from each snow layer is first aggregated and mixed with rainfall on top of the first snow layer:

\[
F_{liq, \text{top}} = I_{sn \cdot Rf} + \int M dz_{sn} \tag{4.10}
\]

\[
R_{F_{liq, \text{top}}} = \frac{I_{sn \cdot Rf} R_{F_{liq, \text{top}}} + \int M R_M dz_{sn}}{F_{liq, \text{top}}} \quad \text{(where } R_M = \alpha^+ R_{sn}) \tag{4.11}
\]

where \( F_{liq} \) in the snow layers is vertical liquid water flux, and \( R_{F_{liq}} \) is the isotopic composition. \( I_{sn \cdot Rf} \) and \( M \) in each layer are liquid input flux for the snow reservoir and snow-melting flux, and \( R_{I_{sn \cdot Rf}} \) and \( R_M \) are their isotopic ratio, respectively; \( \alpha^+ \) is the equilibrium fractionation factor between ice and liquid [Majoube, 1971b], \( R_{sn} \) is the isotopic ratio of snow, and \( z_{sn} \) is the vertical axis for the snow layers.

Liquid water that percolates downward is affected by equilibrium-fractionated refreezing that depends on the freezing point at each layer in the soil:

\[
\frac{\partial F_{liq}}{\partial z_{sn}} = -F_r \tag{4.12}
\]

\[
\frac{\partial R_{F_{liq}} F_{liq}}{\partial z_{sn}} = -F_r R_{F_r} \quad \text{(where } R_{F_r} = R_{F_{liq}}/\alpha^+) \tag{4.13}
\]

Here \( F_r \) and \( R_{F_r} \) are the refreezing flux of liquid water at each layer and the isotopic ratio, respectively. The following equation can therefore be introduced:

\[
\frac{\partial R_{F_{liq}}}{\partial z_{sn}} = (1 - \frac{1}{\alpha^+}) F_r R_{F_{liq}} \tag{4.14}
\]
SWE and its isotopic composition at each snow layer \((S_n \text{ and } R_{sn})\) are given as follows:

\[
\frac{\partial S_n}{\partial t} = -M + F_r \tag{4.15}
\]
\[
\frac{\partial S_n R_{sn}}{\partial t} = -M \alpha^+ R_{sn} + F_r R_{Fr} \tag{4.16}
\]

Therefore,

\[
\frac{\partial R_{sn}}{\partial t} = \frac{M}{S_n} (1 - \alpha^+) R_{sn} + \frac{F_r}{S_n} \left( \frac{R_{Fr}}{\alpha^+ R_{sn}} - 1 \right) \frac{R_{sn}}{\alpha} \tag{4.17}
\]

Note that sublimation from (4.8) and (4.9) and snowfall should also be considered at the top layer as

\[
\frac{\partial S_n}{\partial t}_{\text{top}} = I_{sn} - E_{sn} - M - F_r \tag{4.18}
\]
\[
\frac{\partial S_n R_{sn}}{\partial t}_{\text{top}} = I_{sn} R_{I_{sn}} - E_{sn} R_{E} - M \alpha^+ R_{sn} + F_r R_{Fr} \tag{4.19}
\]

where \(I_{sn} \) and \(R_{I_{sn}} \) are the solid input flux on snow (usually snowfall) and its isotopic ratio.

### 4.2.5 Vegetation

Transpiration flux \(E_t\) estimates include the isoflux that considers equilibrium and kinetic fractionation from liquid to gas at stomata:

\[
E_t = (1 - f_w) q^s C_{E_c} \left[ V_{tl} \left( q^s(T_c) - q_a \right) \right] \left( q^s(T_c) - q_a \right) > 0 \tag{4.20}
\]

\[
R_{E_t} = \alpha_k \left( q^s(T_c) - q_a \right) R_{I_{tl}} \frac{R_{I_{tl}} - q_a R_{I_{tl}}}{q^s(T_c) - q_a} \tag{4.21}
\]

where \(f_w\) is the wet area rate of canopy, and \(C_{E_c}\) is the bulk coefficient for transpiration that depends on Simple Biosphere Model-type (SiB-type) stomata resistance [Sellers et al., 1996] and \(LAI\). \(T_c\) is canopy temperature, and \(R_{I_{tl}}\) is the isotopic ratio in the leaf tip. Equilibrium dew condensation onto the canopy is assumed when \((q^s(T_c) - q_a)\) is negative.

Vegetation in the model has one trunk layer and five leaf layers by default. Transpiring water drawn up from the root zone layers has isotopic composition \(R_t\) that is weighted by the root density in each root zone layer \((f_{root})\) as

\[
R_t = \int \frac{R_{th} f_{root}}{f_{root}} dz \tag{4.22}
\]

where \(R_{th}\) is the isotopic ratio of soil water in each layer, and \(z\) is the vertical axis for soil.

Water conveyance driven by both transpiration and back diffusion is considered in the leaf. The isotopic ratio of leaf water, \(R_t\), can therefore be described as

\[
\frac{dV_l}{dt} \frac{\partial R_t}{\partial z_l} = \frac{1}{\rho} \left[ \left( E_t A_l \frac{\partial R_t}{\partial z_l} - D_r \frac{\partial^2 R_t}{\partial z_l^2} \right) \right] \tag{4.23}
\]

Here, \(z_l\) is the axis directed from leaf base to tip, and \(V_l\) and \(A_l\) are water volume in a leaf and area of leaf surface, respectively, determined by vegetation type. \(LAI\) is the leaf area index, \(D_r\)
is liquid diffusivity of an isotope, which diffusivity varies by temperature and isotopic species, and tau describes the crookedness of the leaf. REs and Rr are used as boundary conditions for the leaf tip and leaf base, respectively. Water conveyance flux Et does not vary on the zl- axis (from leaf base to leaf tip): leaf water volume $V_l$ does not change with time.

In ISOLSM [Riley et al., 2002], and in other isotopic land surface models [e.g., Braud et al., 2005], the isotopic compositions in leaves are analytically diagnosed by assuming very small leaf reservoirs (Equation 20 in Riley et al. [2002]) and isotopic steady state conditions ($R_e = R_E$). In contrast, the present model predicts leaf water content and isotopes without assuming a steady state.

4.2.6 Runoff

The four runoff types in the original MATSIRO (saturation excess runoff [Dunne flow], infiltration excess runoff [Horton flow], base flow, and overflow of the uppermost soil layer) are also considered in Iso-MATSIRO. All water fluxes onto the ground surface are aggregated and some portion of this water, which depends on surface soil conditions and physical soil properties, becomes Dunne flow or Horton flow. These two surface runoff types are therefore isotopically “direct.” Base flow is calculated using the simplified TOPMODEL scheme. Base flow originates from the layer where the water table depth is defined, and the isotopic composition of the base flow is same as that of the soil layer. Finally, overflow has the same isotopic composition as soil water in the uppermost soil layer.

$$R_{O_s} = R_{O_s} = R_I$$

(4.23)

$$R_{O_l} = R_{O_l} = R_{O_l, zl}$$

(4.24)

$$R_{O_o} = R_{O_o} = R_{O_o, opp}$$

(4.25)

$R_{O_s}, R_{O_l}, R_{O_l},$ and $R_{O_o}$ denote isotopic ratios of Dunne flow ($O_s$), Horton flow ($O_h$), the base flow ($O_l$), and the overflow ($O_o$), respectively. $R_I$ is the isotopic ratio of liquid input flux onto the soil surface, and is isotopic ratio of soil moisture. $z_l$ is the water table depth.

4.2.7 Soil

Soil Evaporation

Water does not evaporate from snow-covered soil. Soil resistance and humidity in the surface soil is calculated only in the snow-free part of the domain. If soil is partly or completely frozen, sublimation from ice is considered. The model assumes that surface humidity results from equilibrium fractionation of soil water contents. Soil evaporation flux, $E_s$, and its isotopic ratio, $R_E$, which considers kinetic isotopic fractionation (diffusion to atmosphere), are

$$E_s = (1 - A_{soil})(1 - f_{ice}) \rho C_E s |V_a| \left( h_{soil} q^*_{(T_s)} - q_a \right)$$

(4.26)

$$R_E = \begin{cases} \frac{h_{soil} q^*_{(T_s)} \alpha(T_s) R_{R_{soil, opp}} - q_a R}{R_a / \alpha(T_a)} & E_s > 0 \\ R_a / \alpha(T_a) & E_s \leq 0 \end{cases}$$

(4.27)
\( A_{sn} \) and \( f_{ice} \) are the snow-covered area rate and frozen soil area rate, respectively. \( \widetilde{C}_{Es} \) is a bulk coefficient that considers soil resistance for cases of evaporation \((h_{soil} q^s_{(T_s)} - q_a > 0)\) but not for cases of dew condensation \((h_{soil} q^s_{(T_s)} - q_a \leq 0)\); \( h_{soil} \) is the relative humidity in the surface soil, and \( q^s \) is saturated specific humidity.

Similarly, sublimation flux from frozen soil \( E_{s,ice} \) and its isotopic ratio \( R_{E_{s,ice}} \) are

\[
E_{s,ice} = (1 - A_{sn}) f_{ice} \rho \widetilde{C}_{Es} \lvert V_a \rvert (h_{soil} q^s_{(T_s)} - q_a)
\]

\[
R_{E_{s,ice}} = \begin{cases} 
\alpha_h \frac{h_{soil} q^s_{(T_s)} - q_a}{h_{soil} q^s_{(T_s)} - q_a} & E_s > 0 \\
\frac{R_a}{\alpha_{ice} (T_a)} & E_s \leq 0 
\end{cases}
\]

where \( \alpha_{ice} \) and \( R_{\theta_{top}} \) are respectively the equilibrium sublimation factor (Moujoube, 1971) and isotopic ratio of frozen ice in the surface soil.

### Soil water transfer

Thermal and water transfers occur at five layers in the model; these layers are 5, 20, 75, 100, and 200[cm] (total 4[m]) from the surface. The original MATSIRO uses Richards' equation and thermal conduction to model soil water transfer between the five soil layers with Clapp and Hornberger's [1979] hydraulic conductivity and matric potential. Vapor in soil (and its diffusivity) is ignored. Isotopes in Iso-MATSIRO move only through vertical soil moisture fluxes and source/sink fluxes (i.e., root uptake and runoff). Soil water content and isotopes can therefore be expressed as

\[
\frac{\partial \theta_w}{\partial t} = \frac{\partial F_w}{\partial z} + S_w
\]

\[
\frac{\partial \theta_w R_{\theta_w}}{\partial t} = \frac{\partial F_w R_{F_w}}{\partial z} + S_w R_{\theta_w}
\]

where \( \theta_w \) is soil moisture content, and \( R_{\theta_w} \) is its isotopic ratio; \( F_w \) is the vertical soil moisture flux, and \( R_{F_w} \) is its isotopic ratio; and \( S_w \) is the source term. The flux terms \( F_w \) and \( R_{F_w} \) are expressed using boundary conditions at the surface and the bottom of the soil layers:

\[
F_w = K \left( \frac{\partial \psi}{\partial z} - 1 \right)
\]

\[
R_{F_w} = R_{\theta_w}
\]

at the surface,

\[
F_{w, top} = I_s - E_s
\]

\[
R_{F_{w, top}} = \frac{I_s R_{I_s} - E_s R_{E_s}}{F_{w, top}}
\]

at the bottom,

\[
F_{w, bot} = 0
\]

\( I_s \) is the liquid input flux to the soil surface, and \( R_{I_s} \) is its isotopic ratio. Equilibrium fractionation is considered when soil moisture freezes or soil ice melts.
4. DEVELOPMENT OF ISO-MATSIRO

4.3 One-dimensional offline simulations

4.3.1 Design

One-dimensional simulations in the vertical are based on the iPILPS-EQY1 framework [Henderson-Sellers, 2005], which compares the performance of different land surface models that incorporate isotopes. Simulations for steady states of energy, water, and isotopic fluxes occur at three sites: Tumbaramba, Australia (35S 148E), Manaus, Brazil (3N 60W), and Munich Germany (48N 11E).

Surface boundary conditions, i.e., specific humidity, wind speed, air temperature, pressure, downward radiation, precipitation (convective and large scale), and the isotopic compositions of precipitation (convective and large scale) and vapor, are derived from output from REMO-iso [Sturm et al., 2005] at 15-minute time steps for one ideal year (360 days). The forcing meteorology is therefore modeled and the simulations allow a check on the basic model performance that would be difficult to achieve with observations. Continuous isotopic ratio data for atmospheric vapor and precipitation are very hard to acquire. Table 1 shows selected respective soil types and vegetation types as well as other important parameters.

The model timestep is five minutes, and the model iterates a one-year calculation until differences between the initial and the final water storage (soil, canopy, and snow) decrease below a threshold. Thresholds in this study are 0.01 [mm/year] for water and 0.01 [mm/year*R_{SMOW}] for isotopic species. \( R_{SMOW} \) is the isotopic ratio of SMOW (the standard mean ocean water).

Table 4.1: Soil and vegetation types and corresponding site-specific parameters for three simulated sites

<table>
<thead>
<tr>
<th>soil type(^a)</th>
<th>Tumbaramba</th>
<th>Manaus</th>
<th>Munich</th>
</tr>
</thead>
<tbody>
<tr>
<td>veg. type</td>
<td>mixed coniferous and broadleaf evergreen forest</td>
<td>high latitude deciduous forest and woodland</td>
<td>mixed deciduous forest and woodland</td>
</tr>
<tr>
<td>LAI(^b)</td>
<td>0.63</td>
<td>4.51</td>
<td>4.50</td>
</tr>
<tr>
<td>veg. cover (%)</td>
<td>45</td>
<td>90</td>
<td>90</td>
</tr>
<tr>
<td>root dist. (%)(^c)</td>
<td>15, 40, 40, 5, 0</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

\( a \): Classified by Cosby et al. [1984]

\( b \): Average for target area and fixed in time

\( c \): From the top layer to the fifth layer

4.3.2 Results

Annual budget of energy, water and isotopic fluxes

Table 4.2 shows annual total energy, water, and isotopic fluxes. Energy and water partitioning are plausibly simulated. Transpiration and evaporation from canopy interception are less than soil evaporation at Tumbaramba, but greater than soil evaporation at Manaus and Munich. Table 4.1 suggests that this relationship is largely controlled by LAI and vegetation cover.
Runoff is overestimated at Manaus because of high rates of convective precipitation there (89% of total precipitation, versus 48% at Tumbarumba and 33% at Munich). Convective precipitation is assumed to occur in only a portion of the grid cell (10% by default), so the precipitation rate over the area is ten times the convective rainfall rate, which increases the infiltration excess runoff (Horton flow). In addition, the temporal resolution of the forcing data is too fine, considering the main model purpose of global climate simulation. Finer temporal resolutions cause more fluctuations in precipitation rates and therefore more Horton flow. Discussion of adequate spatial and temporal scales in land surface modeling can be found elsewhere [e.g., Ren and Henderson-Sellers, 2005]. Such details, however, lie outside of the scope of this study.

Table 4.2: Annual budget of energy, water, and isotopic fluxes at three sites

<table>
<thead>
<tr>
<th></th>
<th>Tumbarumba</th>
<th>Manaus</th>
<th>Munich</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Energy fluxes [W/m²]</em></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Net Energy</td>
<td>94.7</td>
<td>141.1</td>
<td>59.6</td>
</tr>
<tr>
<td>Latent Heat</td>
<td>80.4</td>
<td>93.7</td>
<td>74.4</td>
</tr>
<tr>
<td>Sensible Heat</td>
<td>13.9</td>
<td>47.3</td>
<td>-16.8</td>
</tr>
<tr>
<td>Ground Heat</td>
<td>0.4</td>
<td>0.02</td>
<td>2.0</td>
</tr>
<tr>
<td><em>Water fluxes [mm]</em></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Precipitation</td>
<td>1368.7</td>
<td>3072.0</td>
<td>1349.8</td>
</tr>
<tr>
<td>Evaporation</td>
<td>999.1</td>
<td>1165.7</td>
<td>900.7</td>
</tr>
<tr>
<td>Transpiration</td>
<td>112.0</td>
<td>333.8</td>
<td>236.4</td>
</tr>
<tr>
<td>Soil Evap.</td>
<td>697.4</td>
<td>181.3</td>
<td>114.0</td>
</tr>
<tr>
<td>Canopy Evap.</td>
<td>189.7</td>
<td>650.5</td>
<td>548.7</td>
</tr>
<tr>
<td>Snow Subl.</td>
<td>-0.02</td>
<td>-</td>
<td>1.7</td>
</tr>
<tr>
<td>Total runoff</td>
<td>369.6</td>
<td>1906.3</td>
<td>448.8</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Isotopic ratio, $\delta^{18}O/\delta D/d$-excess [permil]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation</td>
</tr>
<tr>
<td>Evaporation</td>
</tr>
<tr>
<td>Transpiration</td>
</tr>
<tr>
<td>Soil Evap.</td>
</tr>
<tr>
<td>Canopy Evap.</td>
</tr>
<tr>
<td>Total runoff</td>
</tr>
</tbody>
</table>

Consider the annual averages of isotopic ratios for each flux. The isotopic ratio of the evapotranspiration flux resembles that of the precipitation, in ±1.5‰ range for $\delta^{18}O$. This confirms that the simple assumption used in atmospheric global simulations, that no fractionation occurs during evaporation from land surfaces [e.g., Hoffmann et al., 1998], is plausible on annual time scales. However, the contents (i.e., evaporation from soil, transpiration, and evaporation from canopy) are isotopically different. Transpired flux is more enriched than precipitation at all three sites. In Tumbarumba, particularly, transpired water is about 4.5‰ heavier than precipitation. In contrast, evaporate from soil is depleted at all sites; the difference between soil
evaporation and transpiration is about 99% for $\delta^{18}$O in Munich. Evaporation from the canopy intercept has an isotopic ratio similar to precipitation. It is depleted in heavy isotopes at Tumbarumba and Munich but enriched at Manaus because of seasonal variations of the isotopic ratio in precipitation.

Furthermore, d-excess in evapotranspiration flux exceeds that in precipitation at all sites. The difference between d-excess of transpiration and soil evaporation is large, more than 10%, with smaller values for transpiration fluxes. Distinct differences in isotopic ratios show that partitioning transpiration and soil evaporation using isotopic information is reasonable on this scale.

Runoff flux is enriched in heavy isotopes at Tumbarumba and Munich, but not at Manaus. Evaporative fractionation affects enrichment. The runoff rate is larger at Manaus where there was less enrichment than the other sites. The table also shows slightly depleted values in runoff at Manaus. Such depletion must arise from seasonal variations because runoff should be enriched to some degree if the precipitation is isotopically steady.

**Seasonal variations in soil isotopes**

Figure 4.2 shows seasonal variations of $\delta^{18}$O in soil moisture at each of the five layers and in precipitation for one year at the three sites. Fluctuations in $\delta^{18}$O decrease with increasing soil depth at all three sites. When precipitation occurs, the first soil layer responds. That layer subsequently becomes enriched as evaporation occurs. The precipitation and evaporation spikes are sharpest at Tumbarumba because soil evaporation dominates the total evapotranspiration there (see Table 4.1). Furthermore, the loamy soil at Tumbarumba is easier to infiltrate because the hydraulic conductivity is larger; thus the first layer of soil will be more sensitive to isotopic variations of precipitation there. Snow exists at Munich from the end of November until April. Therefore, direct soil infiltration of precipitation is prevented, and large fluctuations do not occur there during winter.

Figure 4.3 shows seasonal variations of soil wetness and precipitation. There is less soil wetness at lower levels at Manaus than at the other two sites, even though there is more precipitation. The higher rate of convective precipitation, described in the previous section, means less infiltration to the soil. At Tumbarumba and Munich, there are saturated soil layers for a part of the year.

**Diurnal isotopic cycles**

Figure 4.4 shows diurnal variations of transpiration flux (Figs. 4.4a, 4.4d, and 4.4g), $\delta^{18}$O values (Figs. 4.4b, 4.4e, and 4.4h), and $\delta^{18}$O in leaf and stem water (Figs. 4.4c, 4.4f, and 4.4i). Transpiration at Tumbarumba is smallest because of smaller LAI and smaller vegetation cover rates. Dry root soil layers and a consequent increase in soil resistance cause relatively low transpiration at Manaus in June–August (JJA). Soil dryness is highly sensitive to the soil properties (soil type). Selective root uptake also might be considered for the case of dry soil conditions.

Fluctuations in isotopic ratios in transpired fluxes have a large range, but nocturnal $\delta$-values have little impact of the total flux because the corresponding water flux is very small. During
4. DEVELOPMENT OF ISO-MATSIRO

Figure 4.2: Seasonal variations of daily-averaged isotopic ratios ($\delta^{18}O$) in soil moisture at each layer (lines), and that in precipitation (crosses) at Tumbarumba (a), Manaus (b), and Munich (c) the day, the isotopic ratio rapidly decreases from morning until noon, and increases in the afternoon. The model does not assume a steady state of transpired isotopic flux, so the diurnal fluctuations can be simulated.

The well-known daytime enrichment in a leaf is plausibly simulated, but the degree of maximum enrichment is less than observed [e.g., Gan et al., 2002], partly because the values shown are seasonal averages. If meteorological or hydrological causes lead to stomata closure during the day, little diurnal variation in the isotopic ratio of a leaf can occur. However, the maximum isotopic difference between stem and leaf tip is about 10‰ at all three sites and is still smaller than observed averages. It is possible that the Merlivat and Jouzel [1979] simplification of the kinetic fractionation factors underestimates kinetic enrichment when wind speeds are moderate.
4. DEVELOPMENT OF ISO-MATSIRO

Figure 4.3: Seasonal variations of daily-averaged soil moisture at each layer (lines), and that of precipitation (bars) at Tumbarumba (a), Manaus (b), and Munich (c)

δ-diagram for reservoirs and fluxes

Figure 4.5 shows monthly averaged δD–δ18O relationships of predicted isotopic ratios for representative reservoirs and fluxes. Reservoirs include water intercepted by the canopy, the first soil layer, SWE, and leaf water. Fluxes are precipitation, evapotranspiration (separated into soil evaporation, transpiration, and evaporation from canopy), total runoff, and snowmelt.

The δ-values vary widely in time and by kind and are particularly depleted in soil evaporated flux but enriched in soil water and leaf water. Regardless of the large range in values, however, δ-excesses are mostly uniform at all three sites. However, soil water and leaf water have a slightly lower slope than other variables, which are on the meteoric water line (MWL). The first soil layer and leaf were likely more affected by kinetic fractionation.
Figure 4.4: Diurnal variations of transpiration flux (a, d, g), $\delta^{18}O$ in transpiration flux (b, e, h), and $\delta^{18}O$ in leaf (c, f, i) at Tumbarumba (a–c), Manaus (d–f), and Munich (g–i). Each of four lines shows different seasons, and lines with dots in the right column figures (c, f, i) show $\delta^{18}O$ in stem water.

4.4 A global offline simulation with Iso-MATSIRO and ICM

4.4.1 Design of the coupled simulation

A global simulation coupled Iso-MATSIRO with an atmospheric isotope model. However, unlike an ordinary coupled global simulation, which uses LSM and AGCM, the present study uses the Rayleigh-type isotope circulation model (ICM) from Yoshimura et al. [2003, hereafter Y03] to compute the atmospheric circulation of isotopes. The ICM is forced by external meteorological factors from a reanalysis dataset. Those factors include precipitation, evaporation, precipitable water vapor, and vertically integrated vapor fluxes. The ICM using reanalysis fields yields a “realistic” spatial and temporal distribution of precipitation isotopes and surface vapor isotopes [Yoshimura et al., 2004a]; these isotopes are required variables for Iso-MATSIRO. This experiment compares the land effect of precipitation isotopes to Y03, in which constant
Figure 4.5: Seasonally averaged δD–δ¹⁸O relationship for modeled reservoirs (closed symbols) and fluxes (open symbols) at Tumbarumba (a), Manaus (b), and Munich (c). Size of each symbol shows relative magnitude of corresponding water reservoirs or fluxes. The gray line indicates the meteoric water line (δD = 8 × δ¹⁸O + 10).

This experiment has a horizontal resolution of 1° × 1° and a coupling frequency of 10 minutes. Isotopes in evaporative flux over land surfaces are predicted by Iso-MATSIRO and given to the ICM; isotopes in precipitation flux and ambient vapor are predicted by the ICM and given to Iso-MATSIRO. Figure 4.6 shows the coupling structure. Other meteorological variables are derived from Global Energy and Water Cycle, Asian Monsoon Experiment (GAME) reanalysis data [Yamazaki et al., 2000] every six hours. Soil and vegetation types and other physical parameters are taken from the default setting of the Global Soil Wetness Project 2 [Dirmeyer et al., 2002]. Over oceans, δ¹⁸O in evaporative flux has a fixed value of -5‰, versus -9.4‰ in Y03. The global simulation proceeds from 1 April to 31 October in 1998, after a two-April spin-up. A longer spin-up allows for isotopically stable reservoirs with large quantities of water. However, the GAME reanalysis used in Y03, cover only six months, which is not long enough
4. DEVELOPMENT OF ISO-MATSIRO

for a true spin-up. Only $\delta^{18}$O is simulated in this experiment.

Figure 4.6: Schematic representation of Iso-MATSIRO and ICM coupling.

4.4.2 Results and Validations

Figures 4.7a–4.7e show 7-month averages of global $\delta^{18}$O distributions for principal surface water variables: precipitation (Figure 4.7a), evapotranspiration (Figure 4.7b), runoff (Figure 4.7c), soil storage in the surface layer (Figure 4.7d), and canopy intercepted water (Figure 4.7e). Figure 4.7a shows a large scale “continental effect” in simulated precipitation stretching northeast over Eurasia, north over North America, and west over South America. Precipitation $\delta^{18}$O have a large impact on other reservoirs or fluxes, because they show similar isotopic distributions. However, the influences of complex land surface processes modulate the intensity of the direct impact of precipitation. For example, canopy water (Figure 4.7e) and surface soil storage (Figure 4.7d) are more influenced by precipitation than evaporation (Figure 4.7b) and runoff (Figures 4.7c).

Results suggest that previous treatments of isotopes in land surface models, such as “no fractionation and direct return of precipitation isotopes” [Jouzel et al., 1987; Hoffmann et al., 1998; Mathieu et al., 2002] or “uniform isotopic value in evaporation” [Yoshimura et al., 2003], are too simplistic. Figure 4.8 shows the difference between simulated precipitation with (this study) and without (Y03) the land surface treatment through a comparison to monthly observations by the Global Network of Isotopes in Precipitation (GNIP; IAEA/WMO). The global average of systematic bias decreases from -3.4%/o to 0.9%/o, and the RMSE decreases from 4.6%/o to 3.3%/o. The impact of land surface processes on precipitation isotopes cannot be neglected, and the physical treatment in this study is more reasonable as shown in the better match with observations. Moreover, physical land surface treatments considering isotope circulations are necessary for runoff and river water isotope estimates to yield more reasonable precipitation isotope estimates.
Figure 4.7: Global distribution of weighted average $\delta^{18}O$ of precipitation (a), evapotranspiration flux (b), runoff (c), soil moisture at the first layer (d), and canopy intercepted reservoir (e) from April to October 1998.
Figure 4.8: Simulated monthly precipitation $\delta^{18}O$ are compared with GNIP observations for April to October 1998. Land–atmosphere coupled simulation results are shown as circles, and control simulation results in Y03 (without land surface processes) are shown as crosses.
4.5 Summary and Conclusions

This study developed a land surface model, Iso-MATSIRO, which considers stable water isotopes. The model simulates physically reasonable isotopic fluxes and reservoirs at ground surfaces. The model calculates kinetic and equilibrium fractionation of HDO and H$_2^{18}$O between ice, liquid, and vapor phases, with separate considerations for soil surface evaporation, vegetation transpiration, evaporation from the canopy-intercepted reservoir, and snow sublimation. The snow submodel considers isotopic fractionation in melting of snow, freezing of rain, and re-freezing of melted water. Diffusive isotopic movement (so-called back diffusion) in stem and leaf in addition to conveyance by transpirative flux is considered in the vegetation submodel without assuming a steady state condition. The soil submodel has vertical isotopic transfer in five soil layers, including freezing and melting of soil moisture.

A one-dimensional (vertical) simulation with modeled meteorological forcings was conducted for three sites (Tumbarumba, Manaus, and Munich) following the iPILPS EQY1 framework. Results cannot be compared to observations, but they show plausible and reasonable features in the annual isotopic budget, the seasonal variations of $\delta^{18}$O in soil moisture, the diurnal variations of leaf water with enrichment to some extent, and $\delta$-diagram of representative surface reservoirs and fluxes.

A global simulation used Iso-MATSIRO coupled with an atmospheric ICM. Isotopes in precipitation and ambient vapor were simulated in the ICM and given to Iso-MATSIRO, and isotopes in evapotranspiration were simulated in Iso-MATSIRO and given to the ICM. The coupled run simulated six months in 1998 using the GAME reanalysis. Results were compared with GNIP observations. Simulated precipitation $\delta^{18}$O was closer to observations when Iso-MATSIRO was used.
Chapter 5

Atmosphere-Land and Sea surface coupled simulations

Abstract:
The purpose of this study is to suggest some quantitative interpretation of precipitation isotopes for further understanding of the global hydrologic cycles. To achieve this purpose, two different experiments have been implemented. One with constant field of evaporative isotopic ratio, and one with variable evaporative isotope fields which are the output from Iso-AGCM. The results showed, over the zonal regions around 30 to 60 degrees in both hemispheres, isotopes in evaporation have influences to the precipitation isotopes. Then, isotopically physical processes on water surfaces and land surfaces are incorporated to the reanalysis-forced isotope circulation simulation. The 15-year simulation results show great improvement on the target regions, moreover, the systematic underestimation of precipitation isotopes over low latitudinal regions were also dissolved.

5.1 Introduction

Stable isotopes of water, D and $^{18}$O, are good tracers of the global hydrologic cycles. Climatology of observed precipitation isotopes in GNP (Global Network of Isotopes in Precipitation; IAEA/WMO) are intensively investigated and it is stated that the annual averaged values of precipitation isotopes reasonably fit on regression of two explanation variables, i.e., latitude and altitude [Bowen and Wilkinson, 2002; Bowen and Revenaugh, 2003]. Further, Yoshimura et al. [2003] and Yoshimura et al. [2004a] argued that moisture transport system at large scales is the main control of isotopic variability of precipitation. Likewise, explanation of the isotopic variability has been implemented by many literature and resulted some important findings. However, interpretation “only from” the isotopic values themselves have not been well investigated.

The purpose of this study thus is to suggest some quantitative interpretation tools for precipitation isotopes for further understanding of the global hydrologic cycles. To achieve this purpose, the authors first examine the model performance of the isotope circulation model. By
using the atmospheric circulation fields from a more physically sophisticated model, an isotope-
incorporated general circulation model (ECHAM-is0), as the forcing variables to simple isotope
circulation model, the authors try to investigate the controlling factor on precipitation isotopes.
Moreover, the authors try to extract the regions where the variations of evaporative isotopes
are influential.

5.2 ICM run using isotope-AGCM output —Extraction of re-
regions where impact of evaporative isotopes are substantial

5.2.1 Description

In this section, atmospheric water circulation fields from an isotope-AGCM, ECHAM-is0, are
forced to ICM. The isotopic results (precipitation isotopes) will compared with corresponding
isotope-GCM output, which had been already implemented. Required meteorologic forcings are

- Precipitable water $W$
- Vertically integrated moisture flux vector $\vec{Q}$
- Precipitation $P$
- Evaporation $E$

Particularly above two are re-calculated with three-dimensional data. Then, for an additional
experiment (described below) and the comparison, following isotopic variables are required.

- Isotopes in precipitation $R_P$
- Isotopes in evaporation $R_E$

The spatial resolution of the ECHAM-is0 run was L19T42. With enough spin-up period, 6-
hourly data are obtained for an ideal year. The data are converted to appropriate manner for
ICM ($2.5^\circ \times 2.5^\circ$). The plausibility of basic AGCM performance is evaluated by some previous
studies, i.e., Hoffmann et al. [1998] and Hoffmann and Heimann [1997].

We set two experiments with different manipulation of isotopes in evaporative flux, such as
giving constant values (same as the control run in Yoshimura et al. [2003]) and using given
values by ECHAM-is0. The purpose of the experiments is, at first, to examine the mechanism of
spatio-temporal variability on precipitation isotopes by comparing a simple but robust method
(ICM) and a physically sophisticated method (ECHAM-is0), and to investigate the impact of
variation of evaporative isotopes to the atmosphere, and occasionally, precipitation isotopes.
Schematic design of the experiments are shown in Figure 5.1.

5.2.2 Results

Figure 5.2 shows global distribution of correlation coefficients between daily precipitation
$\delta^{18}O$ variations in our simulated results and the ECHAM-is0 results. In addition to tropical
maritime regions over Indian ocean and Pacific ocean, Indochina Peninsula, northern part of
South America, and the central America, show quite similar daily variation to the ECHAM-
iso simulated results. Even more interestingly, over the eastern part of Eurasian continents
(Siberia), there shows relatively high synchronized variations comparing to other high latitude regions. On the other hand, low correlations are found over extremely arid zone over Sahel and the Middle east, some areas in the central Asia, and over some maritime regions near west coasts i.e., in the subtropical zones over South and North Atlantic ocean, and South Pacific ocean. Near the poles, limited information can be obtained because of the model’s shortcomings (described in Appendix D).

Results from the second experiment are shown in Figure 5.3. The global tendency is similar to that of the first experiment, but slight positive increases are obviously detectable (became more alike to that of ECHAM-iso results). Figure 5.4 shows the difference between the correlations of the two experiments. The large positive differences over zonal regions around 30 to 60 degrees in both hemispheres imply that spatial and temporal heterogeneity of isotopes in evaporation have remarkable influences to the precipitation isotopes. In other words, appropriate modeling or estimates of these heterogeneity should be incorporated for further better understanding of the precipitation isotope signals over middle to high latitudinal regions.

On the other aspect, it is arguable to note that, over the tropical zones, daily variability of precipitation isotopes are more reasonably reproducible regardless of physically reasonable treatments of surface processes for evaporative isotopes. More robust and reliable circulation field (precipitation, humidity, and moisture fluxes) are required to obtain more realistic variations of precipitation isotopes. At current stage, the AGCM’s uncertainty diverges in time even if there is perfect boundary conditions [Koster et al., 2004]. Thus the memory of initial conditions lasts only several days over the low latitudinal zone [Yamada et al., 2005]. In reanalyses, these shortcoming (or just characteristics) of AGCMs are overcome by assimilation with observations, and realistic (or “actual”) circulation fields are simulated as very short-term
Figure 5.2: Result from the first experiment (a run with constant evaporative isotopes): Global distribution of correlation coefficients between daily precipitation $\delta^{18}O$ variations with results from ICM run with ECHAM-isolo circulation fields and the ECHAM-isolo results. Spatio-temporal variations in evaporative isotopes are not taken into account in the ICM run.
Figure 5.3: Result from the second experiment (a run with forcing evaporative isotopes): Global distribution of correlation coefficients between daily precipitation \( \delta^{18}O \) variations with results from ICM run with ECHAM-iso circulation fields and the ECHAM-iso results.

Forecast. The good agreement with precipitation isotopes over low latitude has occasionally been achieved in Yoshimura et al. [2003, 2004a]. In other words, what precipitation isotopes tell us is the reality of atmospheric circulation fields, not that of surface conditions or of meteorological condition “at” the site. These arguments above do not contradict to the conclusion of experiments in Yoshimura et al. [2004a].

Furthermore, the regions where the correlations are low in both experiments indicate that indescribable processes in the simple ICM that are incorporated in isotopic AGCM have large impact on the precipitation isotopes. These regions are found over arid regions including Sahel and the Middle East, and eastern parts of southern Pacific ocean and Atlantic ocean. In these regions, the correlations between daily variations of the second experiment and ECHAM-iso even decreased. It should be noted that Japan is the place with decrease of correlations, too. It is arguably said that the influential processes are the post-condensation processes inside of the clouds and below the clouds. Therefore, for simulations of precipitation isotopes over those regions, the use of AGCM should worth to parameterize or simply explain the actual variations of precipitation isotopes.
5.3 Reanalysis-forced ICM simulations with Craig and Gordon mechanism on the water surfaces

Previous section revealed that isotopic evaporative processes have large impact on variations in precipitation isotopes over middle to high latitudinal regions. In Yoshimura et al. [2004a], on the other hand, the precipitation isotopes over the regions were well simulated at significant levels by using ECMWF (European Center for Mid Weather Forecast) reanalysis product (ERA15; Gibson et al. [1997]), without considering appropriate isotopic treatments on water and land surfaces. However, there are some systematic discrepancy between observations and simulations. They also pointed out the importance of surface treatments for evaporative isotopes to decrease the errors. In this section, the water surface processes for proper isotopic variability in evaporation are incorporated to ICM, and differences between the simulation with and without the processes will be examined.

5.3.1 Water surface kinetic fractionation

Craig and Gordon [1965] modeled kinetic isotopic fractionation of the isotopic composition with consideration of the surrounding vapor. Turbulent diffusion is assumed to produce no isotopic fractionation, although molecular diffusion does. Evaporative flux $E$ and corresponding isoflux $E^{iso}$ are expressed as

$$E = \rho C_E |V_a| \left( \theta^*(T) - \theta_a \right)$$

(5.1)
\[ E^{\text{so}} = \rho C_{E^{\text{so}}} |V_s| \left( q^*(T) R_{\text{sea}} \alpha(T) - q_a R_a \right) \]  

\[ R_E \equiv \frac{E^{\text{so}}}{E} = \frac{C_{E^{\text{so}}}}{C_E} \left[ \frac{q^*(T) \alpha(T) R_{\text{sea}} - q_a R_a}{q^*(T) - q_a} \right] \]  

where \( R_E \) is the isotopic ratio of flux \( E \); \( \rho \) is concentration of water; \( V_s \) indicates wind speed on the surface; \( q^*(T) \) and \( q_a \) are the saturated specific humidity at temperature \( T \) and the specific humidity of the surrounding vapor, respectively; and \( C_E \) and \( C_{E^{\text{so}}} \) denote bulk coefficients of water and isotopes against water vapor, respectively, where surface roughness resistance is considered. The ratio of bulk coefficients, \( C_{E^{\text{so}}}/C_E \) is the kinetic fractionation coefficient, and Merlivat and Jouzel [1979] approximated as functions of wind speed, \( V_s \). \( R_{\text{sea}} \) and \( R_a \) denote the isotopic composition of water in the sea surface and surrounding vapor, respectively. In this simulation, \( R_a \) is taken from vapor isotope ratio of the one-layered box of ICM. \( T \) is temperature at the reservoir, and \( \alpha \) is equilibrium fractionation factor obtained from Majoube [1971a].

The simulation is executed for 1 January 1979 to 31 December 1993 with 6-hourly and 1°x1° meteorological forcings in ERA15, \( i.e., \) precipitation, vertically integrated moisture flux, precipitable water, and evaporation. In addition to these, the surface conditions are required for the calculation of evaporative isotopes at land surface and sea surface. Temperature, windspeed, humidity, and pressure at surface are obtained from ERA15 in the same manner.

### 5.3.2 Results

For comparison, similar figures in Yoshimura et al. [2004a] are drawn in Figures 5.5 to 5.7. In general, slight increases of correlations with observations over most regions, particularly over Europe, have been realized. The simulation with ERA15 circulation fields has already been noted to have better reproducibility of observation in terms of correlation coefficients, but the current experiment exceeded in further better way, as shown in Figures 5.5 and 5.6. Even more important, the systematic discrepancy between observation and simulation were decreased in most sites as shown in Figure 5.7. Particularly, the differences are dramatic at the maritime sites and the coastal sites. In the northern edge of the North America, however, the systematic overestimation was accelerated.

As some examples, several sites with enough observation data are picked up in Figures 5.8. In Bangkok, there were about 5 % underestimates in the previous simulation, but it became almost none, then the lines fit more adequately. At Hong Kong as well, the underestimation was fixed. At these two sites, degrees of seasonal variations' amplitude are reasonably estimated. However, in Ryori, Japan, the observation does not have clear seasonality as all of the simulations show somehow. At Darwin, as similar as Hong Kong and Bangkok, the underestimation of the lower peak is dissolved. At Midway, just the center of the Pacific Ocean, it shows an interesting feature that there is little seasonality, but that of interannual can be seen. At current stage, the evaporation experiment also could not successfully explain this temporal shift. The simulation, however, is seemingly getting better. Ljubljana, in Slovenia, underestimation of the upper peak of the variation was fixed. The seasonality thus is simulated quite well. In Harare, Zimbabwe, simulated results are similar each other. It indicates that isotopic behavior in evaporative process does not have important influence on this location. Next, Manaus would be a target
for further investigation. In addition to seasonal changes, interannual variations are obviously simulated, but the observed values are far enriched in isotopes. It seemingly relates to vegetation or land processes. Therefore, coupled simulations with isotopes enabled land surface model might give some clue to explain the current discrepancy. At Ottawa, amplitude of seasonality became close to that of the observations, but still the simulation underestimates the higher peak during summer. The simulated variation for Stuttgart became closer to the observation, too.

Figure 5.5: Global distribution of correlation coefficients between GNIP observations and simulated predictions using monthly data: results from the 15-year run with ERA15, but with consideration of isotopic processes on land and sea surface (full-coupled). Circles indicate observational sites with statistically significant correlations at a level exceeding 99%, triangles denote a level exceeding 90%, and crosses show sites where the null hypothesis cannot be rejected at the 90% confidence level. Regions within 1000 km of any site are shaded.

The experiments with isotope-AGCM output implied possibility of improvement on simulations of spatio-temporal distribution of precipitation isotopes over middle to high latitudinal zones. Eventually, the “realistic” simulation with the reanalysis meteorology became improved not only middle to high latitudinal zones but also tropical or maritime regions by dissolving the systematic discrepancy of δ18O in precipitation. The improvement was realized by consideration of physical isotopic processes on the water surfaces. Therefore, it is arguably noted that precipitation isotopes are not really a proxy of only atmospheric processes but also integration of various processes. The experiments in this paper revealed that water surface processes should
Figure 5.6: By the experiment with isotopic land and sea surface processes, number of sites in Western Eurasia (WE), Eastern Eurasia (EE), and other regions categorized by correlation significance levels (more than 99%, 95%, 90%, and the other) between the GNIP observations are shown.

Figure 5.7: As in Figure 5.5, but for mean bias. Solid and dashed circles indicate negative biases (underestimation) and positive biases (overestimation), respectively.
Figure 5.8: Observed monthly variations in precipitation $\delta^{18}O$ (red lines) comparing with the three simulations. Blue lines and plots show simulation with land and sea surfaces isotopic fractionation processes with ERA15 meteorological forcings (full-coupled); green lines and plots show the results of constant evaporative isotope ratios with the same reanalysis (control); and pink shows the results only using sea surface isotopic processes and constant evaporative isotopes from land surfaces (sea). From the top, Bangkok (Thailand), Hong Kong (China), Ryori (Japan), Darwin (Australia), and Midway (USA).
Figure 5.9: As in Figures 5.8, but for Ljubljana (Slovenia), Harare (Zimbabwe), Manaus (Brazil), Ottawa (Canada), and Stuttgard (Germany)
be included into account. Once the variability of precipitation isotopes are plausibly simulated, it promises high-level validity of the water-related processes that generates the isotopic variation in the simulation.

5.4 Summary and Conclusions

The purpose of this study is to suggest some quantitative interpretation tools for precipitation isotopes for further understanding of the global hydrologic cycles. To achieve this purpose, the authors first examine the model performance of the isotope circulation model. By using the atmospheric circulation fields from a more physically sophisticated model, an isotope-incorporated general circulation model (ECHAM-is), as the forcing variables to simple isotope circulation model, the authors try to investigate the controlling factor on precipitation isotopes. Moreover, the authors try to extract the regions where the variations of evaporative isotopes are influential.

Two different experiments have been implemented. One with constant field of evaporative isotopic ratio, and one with variable evaporative isotope fields which are the output from Iso-AGCM. The results were compared with Iso-AGCM’s precipitation isotope field, and indicated that over the most of the tropical regions, daily variability of precipitation isotopes are controlled by large scale moisture transporting system and they are easily reproducible regardless of physically reasonable treatments of surface processes for evaporative isotopes nor the post-condensation processes inside or below the clouds. On the other hand, over the zonal regions around 30 to 60 degrees in both hemispheres, spatial and temporal heterogeneity of isotopes in evaporation have remarkable influences to the precipitation isotopes, and for the better simulation of precipitation isotopes over the regions, appropriate modeling or estimates of the heterogeneity of isotopic evaporation fields are necessary. At last, over arid regions over Sahel and Middle East, and coastal maritime regions, have highly dependent on other physical processes, such as post condensation processes.

Then, isotopically physical processes on the water surfaces are incorporated to the reanalysis-forced isotope circulation simulation. The aim of this simulation is to better simulate the precipitation isotope variations over the middle and high latitudinal regions, in accordance of the previous numerical experiments. For appropriate modeling of isotopes on the water surface, the authors adopt Craig and Gordon’s kinetic fractionation model with Merlivat and Jouzel’s modifications. The simulation was forced with atmospheric fields as well as surface conditions from ERA15. The 15-year simulation results show great improvement on the target regions, but moreover, the systematic underestimation of precipitation isotopes over low latitudinal regions were also dissolved. Therefore, it is arguably said that precipitation isotopes are not a proxy of only atmospheric processes but also integration of various processes including surface processes. Thus, once the variability of precipitation isotopes are plausibly realized, it promises the robustness of the water-related processes that generates the isotopic variations.
Chapter 6

Colored Moisture Analysis
Estimates of Variations in 1998
Asian Monsoon Water Sources

Abstract:
This study investigated the dynamic motion of atmospheric water advection by an analytic method called colored moisture analysis (CMA), that allows for the estimation and visualization of atmospheric moisture advection from specific source regions. The CMA water transport model includes balance equations with the upstream scheme and, uses external meteorological forcings. The forcings were obtained from the Global Energy and Water Cycle Experiment (GEWEX) Asian Monsoon Experiments (GAME) reanalysis. A numerical simulation with 79 global sections was run for April to October 1998. The results clearly showed seasonal variations in advection associated with large-scale circulation fields, particularly a difference between rainy and dry seasons associated with the Asian monsoon. The study also proposes a new definition of southwest Asian monsoon onset and decay, based on the amount of water originating from the Indian Ocean. Earliest onset occurs over southeastern Indochina around 16–25 May. Subsequent onset occurs in India one month later. These results agree with previous studies on the Asian monsoon onset/end. The CMA provides a clearer, more integrated view of temporal and spatial changes in atmospheric circulation fields, particularly Asian monsoon activities, than previous studies that focused only on one or two distinct circulation features, such as precipitation or wind speed. Furthermore, monsoon transition in a specific year, 1998, first became analyzable, whereas the previous studies used climatologies. Furthermore, with assumption of perfect reproducibility of precipitation isotopes, the authors found the “precipitation age” is direct interpretation of precipitation isotopes over tropics, $\xi = -0.236 \times \delta^{18}O + 5.64$ has been introduced.
6. **COLORED MOISTURE ANALYSIS**

6.1 **Introduction**

Where does today’s rain originate? This is a simple question for which no clear answer has been given. Many studies have related this question to precipitation recycling [e.g., *Elahhir and Bras*, 1996; *Burde and Zangvil*, 2001]. However, because direct validations are difficult [Burde and Zangvil, 2001], discussions of the appropriateness of several recycling models have traditionally been restricted to underlying limitations or assumptions. However, the use of stable water isotopes [e.g., *Salati et al.*, 1979; *Gat and Matsui*, 1991], which are theoretically actual tracers influenced directly by atmospheric behavior, is suitable for validating recycling models. Interpretation of isotope analysis for water recycling studies is complicated, because a number of simultaneous processes may affect the isotope data.

Studies that identify water by its origin and incorporate tagged water into general circulation models (GCMS) are a direct way to estimate water recycling, and to answer the question of where rains originate [e.g., *Koster et al.*, 1986; *Numaguti*, 1999; *Bosilovich and Schubert*, 2002]. Indeed, *Bosilovich and Schubert* [2002] determined that their estimates were sufficiently reasonable and could be used as validation of simpler bulk diagnostic estimates from recycling models such as those of *Elahhir and Bras* [1994] and *Brubaker et al.* [1993]. Results from GCMS with tagged water, however, are limited by the accuracy of GCMS.

During the past two decades, many studies have incorporated stable isotopes into GCMS [e.g., *Joussaume et al.*, 1984; *Jouzel et al.*, 1987; *Hoffmann et al.*, 1998; *Mathieu et al.*, 2002; *Noone and Simmonds*, 2002]. Such isotope-GCM studies can be directly validated using observed isotopic records. Good model reproductions of observations support the reliability of tagged-water simulations with isotope-GCMs. However, the good reproduction of observed isotopic compositions in precipitation is limited to monthly averages. Indeed, GCM studies have not reproduced short-term variability in the isotopic composition of precipitation. This observed short-term variability is often greater than seasonal or monthly variability. The failure of GCMS to reproduce short-term variability is likely a result of either coarse spatial resolution [Hoffmann et al. 2000], or errors in the predicted circulation fields [Yoshimura et al. 2003], or both, and is a limitation in studies using GCMS.

Recently, a simple global one-layer isotope circulation model that includes Rayleigh distillation, and uses external meteorological forcings, has reproduced short-term (daily) variability over the subtropics, particularly Thailand [Yoshimura et al. 2003]. Results from that study suggest that large-scale moisture transport (more than 100 km in the horizontal) dominates the control of daily variations in the isotopic composition of precipitation. The reproduction of the isotopic signal suggests that it is reasonable to estimate tagged water by replacing the tracers of water isotopes in the model with tagged water. This also allows a discussion of the daily variation of tagged-water behavior, because the isotopic model reproduces daily isotopic variability.

The present study replaced the isotopic tracers used in the water transport model of Yoshimura et al. [2003] with tagged-water tracers, to investigate how water from specific regions is spatially transported and how transport varies temporally, particularly at daily time scales. The second section introduces the tagged-water transport model, and a visualization scheme. The model and the visualization scheme are referred to as “colored moisture analysis” (CMA). The

6.2 Method of Colored Moisture Analysis

Colored moisture analysis (CMA) combines both a tagged-water transport model and a scheme for visualization of results. The CMA reveals how water from specific regions is transported, and how that water is mixed. Descriptions of the water transport model and the visualization scheme follow.

6.2.1 Description of the Tagged-Water Transport Model

The present tagged-water transport model is consistent with the control run of the Rayleigh-type isotope circulation model in Yoshimura et al. [2003]. The most important and most controversial assumption is that water vapor is completely mixed in each grid at every time step (the “well-mixed” assumption; mathematically described below in the equation 6.3 and 6.4). The basis of this assumption is derived from sufficient convective activities and vertical diffusion caused by sheer vertical. Many previous studies on precipitation recycling estimates used this assumption, but its uncertainty has not been quantified [Bosilovich, 2002].

A GCM experiment in Bosilovich [2002] indicated that the assumption could not be always satisfied because of vertical variations in moisture source ratios depending on the presence of convective processes. Nonetheless, Yoshimura et al. [2003] included an isotopic circulation model with the well-mixed assumption and successfully reproduced isotopic signals in precipitation. They concluded that large-scale moisture transports control the isotopic variability in precipitation on daily time scales. Isotopic components and tagged water are essentially the same in the model transport scheme. Furthermore, main target regions in the current study are Asian monsoon regions where convective precipitation dominantly takes place. Therefore, this study adopts the well-mixed assumption to simulate tagged-water transport.

The model has a global grid (1.25°×1.25° and one vertical layer), and each grid is approximated as a trapezoid. Atmospheric water circulation calculations maintain the atmospheric water balance [Oki et al., 1995] by the upstream scheme in each grid at every time step. The model’s 10-minute time step satisfies the Courant-Friedrichs-Lewy (CFL) condition in most regions. Near the poles, however, the CFL condition is violated because of the short distance between adjacent points. As in Yoshimura et al. [2003], calculations were not performed poleward of 85°N or 85°S, and one tag is assigned there instead. This simplification hardly influence low latitudes, the main target of the present study. The model incorporates external meteorological datasets obtained from reanalyses. Variables include precipitable water (total column water vapor), vertically integrated moisture fluxes (zonal and meridional), precipitation, and evaporation. The 6-hourly variables in GAME reanalysis version 1.5 [Yamazaki et al. 2001] were used for the present study, as in the control simulation of Yoshimura et al. [2003]. The simulation ran from 0000 UTC 1 April 1998 to 1800 UTC 31 October 1998.
For the evaporative source regions, the globe was divided into 79 sections, 57 on water and 22 on land, as shown in Figure 6.1 and Table 6.1. These 79 sections mostly follow geographical borders such as oceans or countries. In addition, one tag was assigned for initial water vapor and another for water near the poles (poleward of 85°N and 85°S). Thus, the experiment had 81 tags. Various water types were assigned to the source regions and water evaporating from any location on Earth was tagged as one of 81 kinds of water.

The atmospheric water balance equation by Oki et al. [1995] is given as

$$\frac{\partial W}{\partial t} = - \nabla \cdot \bar{Q} + (E - P) \tag{6.1}$$

where $W$, $- \nabla \cdot \bar{Q}$, $P$, and $E$ represent precipitable water, horizontal water vapor flux convergence, precipitation and evaporation, respectively. $\bar{Q}$ is the vertically integrated vapor flux vector; its components are zonal and meridional fluxes.

Using the well-mixed assumption, the balance of each different water is

$$\frac{\partial w_n}{\partial t} = - \nabla \cdot \frac{w_n}{W} \bar{Q} + \delta_n E - p_n \tag{6.2}$$

where $w_n$ denotes amount of atmospheric water from region $n$; $N$ denotes a total number of the tags; $\delta_n = 1$ for region $n$ and $\delta_n = 0$ otherwise.

Using reanalyzed meteorological data does not inherently satisfy the atmospheric water balance equation (6.1). Studies that include atmospheric water balance equations must manage this problem. A common and simple way is to have one term in (6.1) include a residual value to satisfy the balance. For example, evaporation $E$ is given as the residual of the balance equation in Brubaker et al. [1993] and Kanae et al. [2001]. Likewise, the precipitation $P$ serves as a residual value in Eltahir and Bras [1994]. However, residuals of the balance equation often acquire unrealistic values [Kanae et al., 2001]. Another way is the use of multiple linear regression [Bosilovich and Schubert, 2001]. This way, however, is computationally expensive, so that it is not very appropriate for global calculation in high temporal resolution. In Yoshimura et al. [2003], nudging the reanalyzed precipitable water at every time step, and conserving its isotopic composition ameliorate, the unclosed nature of equation (6.1). This study therefore follows a similar approach. However, instead of conserving isotopic composition, the ratios of tagged water in the precipitable water are conserved as

$$w_n^* = w_n \times \frac{W^*}{W} \quad (\text{for } n = 1 \sim N) \tag{6.5}$$

where $w_n^*$ is the corrected tagged-water amount from region $n$, $W^*$ is the reanalyzed total precipitable water, and $W$ is the calculated total precipitable water in a discretized form of the equation (6.1).
Figure 6.1: Global map of 79 tagged-water source regions. Letters surrounded by rectangles indicate abbreviated names of each region; the capital letters and following lowercase letters denote areas and direction, respectively. Table 6.1 lists the areas abbreviated by capital letters. Directions (small letters) are denoted as n: north, c: center, and s: south, and w: west, m:middle, and e: east. Tags over lands are shaded.
## 6. Colored Moisture Analysis

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6.2.2 Description of the Visualization Method

The tagged water in the transport model output is color-coded for ease of visualization. A mixture of colors determines the content of tagged water at any time, in any grid. Given that three primary colors comprise color, a matrix with three elements that have the same variation ranges denotes color. Hence, mixed color in any grid is

\[
\begin{bmatrix} R \\ G \\ B \end{bmatrix} = \sum_{n} \begin{bmatrix} r_n \\ g_n \\ b_n \end{bmatrix} \times \frac{w_n}{W}
\]

(6.6)

where \( R, G, \) and \( B \) in the left matrix indicate red, green, and blue, respectively, and \( r_n, g_n, \) and \( b_n \) denote given degrees of red, green, and blue for region \( n \), respectively. A mixture of many different colors cannot create cloud information from tagged-water mixing, because color consists of only three elements. In this study, CMA uses up to twelve distinct colors in addition to black (initial water) and white (the rest of the regions).

6.3 Results

6.3.1 Attenuation of Initial Water

Figure 6.2 shows how the initial water, which is computed from all global vapor content on every first day in April to October 1998, decreases. Fitting the data points with exponential curves reveals a range of e-folding times for atmospheric water between 7.3 days (April) and 9.2 days (August). These values are comparable to Trenberth’s [1998] global mean estimate of 8.4 (April) to 9.1 (August) days (shown by the exponential curve of GAME reanalysis’s April in Figure 6.2) using global monthly averages of precipitation and precipitable water from GAME reanalysis for every month (April to October) 1998, and to Bosilovich et al.’s [2002] estimate of 9.2 days using GEOS-3 (Goddard Earth Observing System version 3) GCM runs initialized in May with 1991 SSTs. This correspondence indicates that the global water budget in GAME reanalysis is appropriately balanced. In addition, it suggests that differences between estimates of the global water budget that do not take moisture transport into account [Trenberth, 1998] and those that do (this study) are likely to be very small.

6.3.2 Continental Cycling Ratio

Figure 6.3 shows the global distribution of the percentage of precipitation that originates as water over land. The average represents May to October 1998. Initial water was removed within one month as shown in Figure 6.2. Thus, this figure excludes data from April which would be influenced by initial water. The ratio of precipitation that originates over land to total amount is termed the “continental cycling ratio.” The degree of water originated from land is important, because it gives useful information on possible interactions between land surface hydrology and climate [Borde and Zangvil, 2001]. The term “recycling ratio” in many previous studies (e.g., Burde and Zangvil [2001] usually suggests that part of the precipitation over a limited area originates from evaporation from that same area (referred to as “local evaporation”). In Figure
Figure 6.2: Attenuation of remaining initial assigned water ratio. Shading indicates range of simulated results for each month (April to October) 1998, assuming that water uniformly covers the whole globe. Diamonds show the April result comparing with Trenberth's [1998] estimation of the same month (solid line), using the same data source (GAME reanalysis).

6.3, however, evaporation from all land surfaces is summed and the percentage of its contribution to precipitation is shown. Because the recycling ratio is sensitive to spatial scales [e.g., Trenberth, 1999], comparison to past studies must include careful consideration of the spatial scale of target regions. When a global continental cycling ratio is calculated, however, the target area is fixed to all land surfaces over the entire globe. Thus, results can be easily compared with similar studies. Figure 6.3 includes geographical features on continental scales and is comparable to a similar figure in Bosilovich et al. [2002] which shows the average of 15-year GEOS-3 GCM runs. It must be recalled that the continental cycling ratio in Figure 6.3 was an average for a half year in 1998. Nonetheless, the similar results imply that the simplified water transport model, with the well-mix assumption, reasonably works for calculation of precipitation sources.

Over the central part of each continent, precipitation contains much continental cycling water. Over Eurasia, for example, over 80% of precipitation originates from land surface evaporation, with extrema around the Tibetan Plateau and the Mongolian High region. Over Africa, regions near the Equator have a very high continental cycling ratio.

6.3.3 Mixing and Transport of Tagged Water: Results of CMA

The experiment partitioned the globe into 79 regions (Figure 6.1). Figure 6.4 shows the color assignments for the visualization of results. The main foci of this experiment are regions influenced by the Asian monsoon. The regionally visualized tagged-water mixing and transport are shown in Figure 6.5. The maps are snapshots of consecutive 10-day periods from April to October 1998.
Figure 6.3: Percentage of precipitation that originated as evaporation from all land surfaces; six-month average from May to October 1998.
Initial water, which was assigned the color black, vanishes by the end of April (Figure 6.5c). The time of disappearance corresponds with the residence time of atmospheric moisture as shown in the previous section. Then, the water from the Bay of Bengal (BB; yellow) overspreads northern Indochina (around northern Thailand). Water from the southern part of the North Pacific (NP-sw; violet) covers the southeast part of Indochina (Vietnam and Cambodia).

After mid-May, water from the Indian Ocean is notable and expands northeastward over the Bay of Bengal and Indochina. On May 21 (Figure 6.5f), water from the Bay of Bengal (yellow) is nearly unseen, which suggests that it has mixed with water of other regions. At the same time, water from the Indian Ocean reaches Indochina. Water originating over the Arabian Sea (ARS; pink) has not mixed with water from other regions.

After June, as the summer Asian monsoon becomes active, a sequential, wavy flow from the southern Indian Ocean to Indochina shows (Figures 6.5h–6.5j). The flow moves westward across the southern Indian Ocean, crosses the Equator near the east coast of Africa (near Somalia), moves over the Arabian Sea, curves around India, passes the Bay of Bengal, and finally reaches Indochina. Note how the summer monsoon activates transport and mixing of tagged water in the regions. Thus, distinct colors such as yellow (Bay of Bengal), or pink (Arabian Sea), are not apparent.

During July and August (Figures 6.5k–6.5p), this wavy flow remains evident and shows meandering motions. In September, however, as the summer monsoon winds down, water originating from Indochina (assigned as white) itself, appears over Indochina (Figure 6.5r). Furthermore, water from China (assigned as blue) expands southward toward Indochina (Figure 6.5s). In October, Indochina again receives water from the Pacific Ocean as well as from China (Figure 6.5t). This southwestward flow marks the end of the summer monsoon. Moreover, as the wavy flow diminishes, distinct colors again become apparent, such as pinks over the Arabian Sea.

### 6.4 Quantitative Analyses of CMA results

Colored moisture analysis shows that water evaporated from one place is transported to other places. Results over regions influenced by the Asian monsoon show distinctive seasonality. Water evaporated from the Indian Ocean overspreads Indochina in early May. This water then retreats at the end of September and is replaced by water from China or the Pacific. The visualized results from the CMA capture dynamic motions of atmospheric water transport. However, these results are qualitative, because human eyes cannot distinguish very small color differences. Moreover, the visualization changes if different colors are assigned to each tag. In this section, results are quantitatively examined, with emphasis on regions influenced by the Asian monsoon, and on the transport of water from the Indian Ocean.

#### 6.4.1 Temporal Evolution of Moisture Origins in Thailand and India

Figure 6.6 shows temporal variations of tagged-water components in precipitable water. Variations are displayed on a grid including (a) Bangkok, Thailand, and (b) Calcutta, India. Note
that some tagged-water components are aggregated into larger parts for make the figures clearer, i.e. Pacific Ocean, Indian Ocean, and all land regions.

These are indeed answers to the first question in the beginning of this chapter. Figure 6.7 shows percentages of contribution from each origin to monthly precipitation amount. Just after the turnover of initial water (beginning of May), at Bangkok, moisture from the Indian ocean was much (34%), while moisture from land surfaces was intensively large at Calcutta (66%). In July, more than half of precipitation was originated from the Indian ocean at Bangkok (67%), whereas increasing Indian ocean water and decreasing land water, became the same level (28% and 36%, respectively) at Calcutta. In August and September, Indian ocean water was the main source of precipitation at both regions. In October, water from all land surfaces and the Pacific ocean became dominant contributors to precipitation at Bangkok; 39% from land surfaces and 29% from the Pacific ocean. At Calcutta, too, land water became dominant again (39%).

6.4.2 An Application of CMA: Onset and End of the Rainy Season in Regions Influenced by the Asian Monsoon

Back to the temporal variation in water origins in Figure 6.6, Indian Ocean water suddenly increases in mid-May (specifically, on 17 May ) and decreases at the end of September at Bangkok. At Calcutta, interestingly, the increase of Indian ocean water occurs in the middle of June, and the decrease is in October. These dates apparently correspond with the onset and end of the monsoonal rainy season in the Asian monsoon regions. Moreover, it is easily conceivable that the water from Indian ocean reaches southeast Asia, because the monsoonal wind fields are
Figure 6.5: Results of CMA. Mixing and transport of tagged water were assigned to 79 regions over the whole globe and focused on the Asian Monsoon regions. Figures a–v correspond to consecutive ten-day snapshots starting 1 April 1998.
Figure 6.5: continued (u and v)

Figure 6.6: Daily time series variation in water origins of total precipitable water at (a) Bangkok (13.8°N, 100.5°E), Thailand, and (b) Calcutta (22.5°N, 88.4°E), India.
from the Indian ocean towards the south and southeast Asia in those months. Therefore, the
temporal variation in water from the Indian ocean can be related with the temporal evolution
of the Asian monsoon.

This phenomenon is a result of dynamic and integrated change derived from the seasonality
of the Asian monsoon, including temporal and spatial changes in wind speed and direction, rain
area, and humidity. Previous studies have suggested several ways to define characteristics of
Asian monsoon activity, including wind speed and direction [e.g., Ramage, 1972; Matsumoto,
1992], outgoing longwave radiations [e.g., Murakami and Matsumoto, 1994; Matsumoto and
Murakami, 2002], rainfall amount [Wang and LinHo, 2002], and cloud amount [Tanaka, 1992].
Each of these elements is connected to or influenced by the other elements. A significant
temporal change in tagged water is another example of integrated monsoon-related elements.

Therefore, the increase and decrease of water evaporated from the Indian Ocean may be used
to define the onset and end of the Asian monsoon. Matsumoto [1997] defines the onset using
precipitation amounts. That is, the first pentad day of three sequential pentad days that exceed the annual mean precipitation during a year (January to December) is defined as the monsoon’s onset, and the last pentad day is defined as the monsoon’s end. Instead of precipitation, this study uses the amount of water from the Indian Ocean in the local precipitable water as another monsoonal index.

Figure 6.8 displays how pentad averaged Indian ocean water in precipitable water increases and decreases as time passes, averaged over the Indochina peninsula and India (IP and IND in Figure 6.1, respectively). By using the new definition, pentad numbers of onset are 33 (10–14 June) and 36 (25–29 June) over respective regions. Likewise, ends are 52 (13–17 September) and 55 (28 September – 2 October). Both the onset and end over India occur later than those over the Indochina peninsula.

These results are spatially averaged, Figure 6.9 thus shows the spatial distribution of the onset and end dates in each grid. Figure 6.9a shows that the earliest onset occurs over southeast Indochina (around southern Thailand and Cambodia) on pentad 28–29 (16–25 May). Onset then proceeds northward to northern Thailand and Myanmar, eventually reaching India on pentad 36–37 (25 June – 4 July). The transition to the monsoon end is similar, but reversed. Figure 6.9b shows that the earliest monsoon end occurs over northern Indochina around pentad 50–51 (3–12 September). The northern Arabian Sea also has an early monsoon end. The monsoon then retreats southward from both regions, almost in parallel with latitudinal lines. Finally, the monsoon end reaches southern Thailand (around Phuket), and Malaysia on pentad 59–60 (18–27 October).

The steady evolution of the monsoon onset and end correspond with results found by Matsumoto [1997] and Wang and LinHo [2002], who used pentad precipitation, and by Tanaka [1992] who used cloud amount. Agreement with previous studies may be merely a matter of course, because the temporal and spatial variations of precipitation are both factors that cause change in tagged water. However, the tagged-water transport used here results from wind and humidity in addition to precipitation. Thus, tagged-water analysis gives an integrated result and understanding of dynamic atmospheric processes, particularly for the Asian monsoon.

The result shown here is for a specific northern summer in 1998, which was characterized as anomalous by several studies including Ding and Liu [2001] and Shen et al. [2001]. Thus the situation is likely different in other years. In this regard, longer simulations and analyses on interannual variations of water origins or of the monsoon evolution would be useful for further understanding of the monsoon system.

Furthermore, the new definition with CMA results here clearly displays the monsoon evolution in an annual basis, whereas all of the previous studies above used climatologies to define the monsoon. Just for comparison, Figure 6.10 shows the monsoon evolution using the similar definition but for traditional pentad averaged water cycle \((P/W)\) obtained from the GAME reanalysis. Figure 6.10 is not as clear as Figure 6.9. It can therefore be said that the traditional definition with data in an annual basis does not work so well in most regions as using CMA results. This is indeed a new understanding obtained by determining the water sources with CMA; nevertheless the analysis related with the Asian monsoon is one example of CMA applications.
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Figure 6.8: Pentad variations in Indian ocean water in precipitable water from May to October, 1998, averaged over Indochina Peninsula (a) and India (b). Dashed lines denote averages of this period. Dots indicate onset and end dates of Asian monsoon by using the new monsoonal definition with Indian ocean water.

6.5 Estimation of precipitation age with water isotopes

6.5.1 Descriptions of Age-CMA

A combined model with ICM (isotope circulation model; Yoshimura et al. [2003]) and CMA (colored moisture analysis; Yoshimura et al. [2004b]) is developed. This global model simulates tagged moisture advection and corresponding isotopic ratio. Total precipitation is the summation of all precipitated tagged water, so as the isotopic ratio of total precipitation is the weighted summation of all of these tags. There is negligible difference in respective numerical results between the combined model with plural tags and the original ICM (with one tag).
Figure 6.9: Transition of the monsoon onset date (a) and end date (b) using Indian Ocean water in the definition of total precipitable water (see manuscript for details). The numbers with contour lines indicate pentad numbers from 1 January. Regions where the average of the Indian Ocean’s water for the computation period exceeds 5 mm are shaded. Gray levels of both figures denote pentads (lighter is earlier). Hatching indicates the Indian Ocean region.

Yoshimura et al. [2004b] stated that there was little relationship between origin of precipitation and precipitation isotopes, but the relationship of large scale hydrologic cycle mainly controls precipitation isotopes [Yoshimura et al., 2003]. In this study, tags are attributed with time, not with space. So the moisture conservation equation for tagged water $w_n$ is expressed as,

$$\frac{\partial w_n}{\partial t} = - \nabla \cdot \left( \frac{w_n}{W} \bar{Q} \right) + \delta_n E - p_n$$

(6.7)
6. COLORED MOISTURE ANALYSIS

(a) Water Cycle (P/W) Onset Date

(b) Water Cycle (P/W) End Date

Figure 6.10: As in Figure 6.9, but for using pentad water cycling ratio (P/W) as monsoon onset/end definition.

\[
\delta_n = \begin{cases} 
1 & \text{during the certain time} \\
0 & \text{otherwise}
\end{cases} 
\]  

(6.8)

where \( \tilde{Q} \) is the moisture flux vector, \( W \) denotes precipitable water, and \( p_n \) is precipitating tagged water. “Well-mixed” condition constrains the simulation as,

\[
W = \sum_{n=1}^{N} u_n \]

(6.9)

\[
p_n = \frac{u_n}{W} P. \]

(6.10)
Furthermore, corresponding isotopic ratio for tagged water, $R_{wu}$ is described as follows:

$$\frac{\partial R_{wu} w_n}{\partial t} = -\nabla \cdot (R_{wu} \frac{w_n}{W} \vec{Q}) + R_E \delta_n E + \alpha R_{wu} p_n \quad (6.11)$$

where $\alpha$ is the isotopic fractionation factor and $R_E$ is isotopic ratio in evaporative flux.

Number of tags ($N$) is 30 in daily basis. It is enough number for replacement of the old water (30 days before) because most of the old tagged water diminishes in 30 days [Yoshimura et al., 2004b]. The averaged age of precipitation is defined as averages of tagged water contents weighted by number of days from evaporation of the tagged water. Therefore,

$$\bar{\xi} = \frac{1}{P} \sum_{n} p_n \xi_n \quad (6.12)$$

where $\xi_n$ passing days since the last evaporation for tag number $n$ ($\xi_n = \text{mod}(T - n, N) + 1$).

6-hourly meteorological variables (precipitation, evaporation, precipitable water, and vertically integrated moisture flux) in ERA15 (ECMWF Reanalysis for 15 years) are used as forcing data for a 1979–1993 run, as same as the previous section.

### 6.5.2 Results

Figures 6.11 show the seasonal variations of simulated precipitation isotopes and precipitation age. It is clear that in Chiangmai and Manaus, the variations of isotopes are obviously synchronized to that of precipitation age, but in Munich, the both variations are individual each other. Therefore in Chiangmai or Manaus, values of precipitation isotopes are interpreted as precipitation age. The linearly fitted relationship is

$$\bar{\xi}_M = -0.245 \times \delta^{18}O + 5.82 \quad \text{[days]} \quad (6.13)$$

$$\bar{\xi}_C = -0.192 \times \delta^{18}O + 7.26 \quad \text{[days]} \quad (6.14)$$

respectively. Figures 6.12 and 6.13 extract the regions where variations on daily values of precipitation isotopes are significantly correlated with variations on age of precipitation. Seasonal correlation, intercept and slope are averaged for 15 years. Daily precipitation $\delta^{18}O$ in shaded regions in these figures can be interpreted to their age according to these information.

### 6.6 Summary and Conclusions

Circulation models based on the atmospheric balance equation with the upstream scheme and external meteorological forcings have computed distributions of tagged water from specific regions. The accuracy of computations has been inferred from the good isotopic reproduction of Rayleigh-type isotope circulation models with the same atmospheric water transport scheme, including the same forcings [Yoshimura et al. 2003]. Moreover, good isotopic reproduction on the daily time scale mirrors the daily variation of computed tagged-water distributions. Thus, this study discusses the temporal transition of the tagged-water distributions on daily scales.

A numerical simulation with 79 global sections was run for one Northern Hemisphere summer: April to October 1998. The initially assigned water was reasonably attenuated. The e-folding
Figure 6.11: Seasonal variations on δ¹⁸O in daily precipitation and corresponding age of precipitation over near Chiangmai (100E, 17N), Thailand (a), near Manaus (60W, 3S), Brazil (b), and Munich (11E, 48N), Germany (c), for 1985.
Figure 6.12: Linear relationship of $\delta^{18}O$ in precipitation and precipitation age for DJF, 15-year average. Where both daily variations are synchronizing is shaded. The panel shows slope of relationship (a), intercept (b), and the square of correlation coefficients ($R^2$) between these variations (c).

Figure 6.13: As in Figures 6.12, but for SON

time of atmospheric water on 1 April was 7.2 days. This time was comparable with bulk estimates by Trenberth [1998] for April 1998 using GAME reanalysis data, and with a tagged-water simulation using GEOS-3 GCM [Bosilovich et al., 2002]. This implies not only that the global water budget was appropriately balanced, but also that moisture transport was reasonably reproduced by the model simulations.

A global map of the percentage of precipitation that originated as evaporation from land surfaces (the continental cycling ratio) is shown. Inland areas of continents have very high rates of continental cycled precipitation. The center of Eurasia, near the Tibetan Plateau and the Mongolian High region, has the highest ratio (exceeding 80%). The distribution corresponds with estimates by Bosilovich et al.’s [2002] 15-year simulation using GEOS-3 GCM data. This
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similarity also gives justification to the simplicity of the water transport model.

Transport and mixing of the tagged water was visualized by the colored moisture analysis (CMA) process. Moisture transport across the Equator, associated with the southwest monsoon from the end of May to the end of September, showed clearly. Although these visualized figures are only conceptual and qualitative because the visualization method is highly dependent on the arbitrary color assignment of each tag, the figures clearly outline sequential changes, especially the distinct temporal changes caused by the Asian Monsoon.

Time sequences of variations in tagged water at Chiangmai and Bangkok, Thailand, were displayed. The results can help answer the question of where today’s rain originates. For example, according to Figure 6.7, at Chiangmai and Bangkok, Thailand, more than 50% of the precipitation came from the Indian Ocean in July 1998. The relationship between the onset and the end of the rainy season was discussed. Water that originated in the Indian Ocean reached Indochina at the middle of May and retreated at the end of September. This period corresponds to the rainy season in the region (May to October). The present results suggest a new definition for the onset and the end of Asian Monsoon activity, namely the increase and decrease of water from the Indian Ocean, which can be computed by tagged-water transport simulations. The evolving process of the onset and end date was shown using this definition. Results agree with previous studies, including those of Matsumoto [1997] and Wang and LinHo [2002] using pentad precipitation, and that of Tanaka [1992] using cloud amounts.

Elements that form the distinct activities of Asian monsoon, such as precipitation, wind speed and direction, and humidity, are closely interrelated. Thus, compared to analyses that rely on single elements, integrated analysis such as tagged-water analysis gives a clearer view of phenomena. As a result, monsoon transition in a specific year first became analyzable with CMA, whereas all of the previous studies used climatologies of those elements. Note that the method’s good isotopic reproduction of Yoshimura et al.’s [2003] daily- and global-scale precipitation estimates validates the method’s effectiveness.

In this study, the Asian monsoon was the main focus as an example of applications of CMA, but the results from CMA covers the whole globe (except regions poleward of 85°), and they can help to understand phenomena in other regional climates. Moreover, the simulation period was only one Northern Hemisphere summer in this study. This is because of data availability in the GAME reanalysis. Thus, longer simulations may be carried out using data for longer period, and their results can investigate annual, and interannual, variation of atmospheric water advection.

Finally, as conclusive finding for the current status on this study, the authors assume that the simulation for precipitation isotopes are perfect, then try to illustrate what can be known only from the precipitation isotopes information. The authors suggested the “precipitation age” for the question. By tagged water simulation, water are tagged by the time of birth (evaporation), then average staying time in the atmosphere until precipitation can be estimated. Because of perfect model, that is validated by precipitation isotopes, this precipitation age is also rigid. Then simulated precipitation isotopes and precipitation ages are synchronously varies at certain regions over mainly tropics. For these regions, simple linear regressions are applied, then some equations such as averaged precipitation age \( \bar{\xi} = -0.25 \times \delta^{18}O + 5.8 \) has been introduced in Manaus. These are just suggestions under the assumption of the perfect model, but these
kinds of practical interpretation of just only precipitation isotopes or those with few other physical information are very important for engineering point of views and should help to further understanding of mechanism of the global hydrologic cycles.
Chapter 7

Conclusion

This thesis, as a study on systems of global hydrologic cycles with the use of stable water isotopes (HDO and H$_2^{18}$O), challenged to describe why this rain has these isotopic values, and what these isotopic values tell us.

The author first creates models that incorporate isotopic processes to quantitatively induce the mechanism of observed variability of precipitation isotopes. The model is designed to adopt additional physical processes as model components that possibly influence the variability, so that author could examine degree of the influences to precipitation isotopes. Concurrently, the author suggests a new use of water isotopes, namely “evaluation of the global hydrologic cycles by precipitation isotopes.” The framework of the evaluation was developed in this thesis. The knowledge from the inductive modeling study above improves the performance of the evaluation, and provides characteristic and interpretation of evaluated results. Furthermore, analytic investigations of the dynamic motion of atmospheric water advection are derivatively performed for finding of more direct interpretation of the isotopology.

The second chapter seeks to explain the causes of short-term (1–10 days) variability in precipitation isotopes, on which two types of models, i.e., Rayleigh-type models and isotope-AGCMs (Atmospheric General Circulation Models), in the past have not been able to examine clearly. A new water isotope circulation model on a global scale that includes a Rayleigh equation and the use of external meteorological forcings is developed. Transport and mixing processes of water masses and isotopes that have been neglected in earlier Rayleigh models are included in the new model. A simulation of $^{18}$O for 1998 is forced with data from the Global Energy and Water Cycle Experiment (GEWEX) Asian Monsoon Experiments (GAME) reanalysis. The results are validated by GNIP (Global network of isotopes in precipitation) monthly observations with correlation $R=0.76$ and a significance level $>99\%$, and daily observations at three sites in Thailand with similar correlation and significance. A quantitative analysis of the results shows that among three factors that cause isotopic variability, the contribution of moisture flux is the largest, accounting for 37% at Chiangmai, and 46% globally. This highlights the importance of transport and mixing of airmasses with different isotopic concentrations. A sensitivity analysis of the temporal and spatial resolution required for each variable is also made, and the model is applied to two additional datasets. The more accurate GPCP (Global Precipitation Climatology Project) precipitation dataset yields improved model results at all three obser-
vation sites in Thailand. the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis allows the simulation to cover two years, reproducing reasonable interannual isotopic variability.

A global Rayleigh-type H$_2^{18}$O circulation model developed in the second chapter was forced by two reanalysis products, NCEP/NCAR reanalysis (NRAI) and the European Centre for Medium-Range Weather Forecasts (ECMWF) 15-year reanalysis (ERA15) datasets, and the GNIP database was used to validate for 1979-1993 on monthly to inter-annual scales, in the third chapter. We find out that the model monthly reproducibility is better with ERA15, particularly in Europe; the number of sites with a statistically >99% significant correlation is 87 out of 116 available European observational sites, whereas 4 sites are qualified with NRAI. Meanwhile, both monthly results are relatively similar and in good agreement with the observations in the rest of regions. Regardless of small number of comparable sites, we examine that inter-annual δ$^{18}$O variability for 15 years is also better reproduced with the ERA15 in Europe, while there is quite similar reproducibility in the rest of regions, too. This study thus reveals that inconsistency of temporal variations in precipitation between the two re-analyses mainly, but indirectly, causes this difference of isotopic reproducibility. The isotope analyses diagnose accuracy of two-dimensional water circulation fields in datasets with a particular focus on precipitation processes, but in a slightly different manner over middle and high-latitude and over low-latitude: severer over former regions.

In the fourth chapter, the author describes Iso-MATSIRO (Minimal Advanced Treatments of Surface Interaction and Runoff), a land surface model that includes stable water isotopes and simulates physically reasonable isotopic fluxes and reservoirs at the ground. The model calculates kinetic and equilibrium fractionation of HDO and H$_2^{18}$O between ice, liquid, and vapor phases and separately considers soil surface evaporation, vegetation transpiration, evaporation from the canopy-intercepted reservoir, and snow sublimation. One-dimensional simulations with modeled meteorological forcings showed plausible features in the annual isotopic budget, seasonal variations of δ$^{18}$O in soil moisture, diurnal variations of leaf water with some enrichment, and δ-diagram of representative surface reservoirs and fluxes. A subsequent, independent global simulation used Iso-MATSIRO coupled with an atmospheric isotope circulation model for a half year in 1998. Simulated precipitation δ$^{18}$O was closer to observations than in a previous study, confirming the physical treatments of isotopes in the land surface processes.

The fifth chapter first implemented two different experiments to compare ICM developed by the author and Iso-AGCMs. One with constant field of evaporative isotopic ratio, and one with variable evaporative isotope fields which are the output from Iso-AGCM. The results showed, over the zonal regions around 30 to 60 degrees in both hemispheres, isotopes in evaporation have influences to the precipitation isotopes. Then, isotopically physical processes on the water surfaces are incorporated to the reanalysis-forced isotope circulation simulation. The 15-year simulation results show great improvement on the target regions, moreover, the systematic underestimation of precipitation isotopes over low latitudinal regions were also dissolved.

In the sixth chapter, the author tries to suggest some quantitative interpretation of precipitation isotopes with more precise and more reasonable global hydrologic cycles derived from the water circulation scheme that have been evaluated by using stable water isotopes in the previous chapters. In this chapter, the author investigated the dynamic motion of atmospheric
water advection by an analytic method called colored moisture analysis (CMA), that allows for the estimation and visualization of atmospheric moisture advection from specific source regions. The CMA water transport model includes balance equations with the upstream scheme and, uses external meteorological forcings. The forcings were obtained from the GAME reanalysis, as the previous chapter. A numerical simulation with 79 global sections was run for April to October 1998. The results clearly showed seasonal variations in advection associated with large-scale circulation fields, particularly a difference between rainy and dry seasons associated with the Asian monsoon. The study also proposes a new definition of southwest Asian monsoon onset and decay, based on the amount of water originating from the Indian Ocean. Earliest onset occurs over southeastern Indochina around 16–25 May. Subsequent onset occurs in India one month later. These results agree with previous studies on the Asian monsoon onset/end. The CMA provides a clearer, more integrated view of temporal and spatial changes in atmospheric circulation fields, particularly Asian monsoon activities, than previous studies that focused only on one or two distinct circulation features, such as precipitation or wind speed. Furthermore, monsoon transition in a specific year, 1998, first became analyzable, whereas the previous studies used climatologies. Finally, with assumption of perfect reproducibility of precipitation isotopes, the authors found the “precipitation age” is direct interpretation of precipitation isotopes over tropics, $\xi = -0.25 \times \delta^{18}O + 5.8$ at Manaus, Brazil, is introduced.
Appendix A

Development of an Isotopic River Flow Scheme with a Bucket-type Land Model

Stable water isotopes (D and $^{18}$O) in river water, which have integrated information of land surface processes, have been firstly reproduced by an isotopic bucket model (Iso-Bucket) and an isotopic river flow scheme (Iso-TRIP) in this study. Isotopic fractionation by three types of evaporation from land, i.e., evaporation, transpiration, and evaporation from intercepted water, are incorporated in Iso-Bucket. Further, a bucket model is assumed for runoff and fractionation calculation.

Authors suggest the use of river water isotopes for validation of LSMs output related with water transport from an aspect of water quality. Moreover, this study enables us to reasonably estimate variability of isotopic values of evapotranspirated water from land surface. That probably leads better reproduction of spatial and temporal distribution of precipitation isotopes with an atmospheric isotope circulation model.

A.1 Development of a Bucket-type Land Model with Water Isotopes

Input data for the model (Iso-Bucket) are precipitation $P$, evapotranspiration $E$, and $\delta^{18}$O in precipitation $\delta p$. In this section, Author used data for 1998-2001, daily, $2.5^\circ \times 2.5^\circ$ on the globe. $P$ and $E$ are taken from the NCEP reanalysis [Kalnay et al., 1996], and $\delta p$ is from the atmospheric isotope circulation model [Yoshimura et al., 2003] using the same reanalysis.

As schematically shown in the left figure in Figure A.1, the Iso-Bucket adopts a simple bucket scheme for vertical one direction. Depth of the bucket is assumed 150 mm on the whole globe. Runoff $R$ occurs only when water exceeds the 150mm-bucket by an equation

$$\Delta S = P - E - R \quad (A.1)$$
where $\Delta S$ is temporal change in storage water in soil (water depth in the bucket, in other words).

In the model, isotopic fractionations by evapotranspiration processes are distinguished by three types, i.e. evaporation from surface soil $E_s$, transpiration from vegetation $E_t$, and evaporation from intercepted water by plant leaves $E_i$. While evaporation from surface soil assumes to fractionate water isotopes by the Rayleigh equation, transpiration and evaporation from intercepted water do not fractionate isotopes. Therefore, following equations are used.

\[
E = E_s + E_t + E_i \quad (A.2)
\]

\[
\delta e_t = \delta r = \delta s_{(t)} \quad (A.3)
\]

\[
\delta e_i = \delta p \quad (A.4)
\]

Rates of each evaporation are taken from JMA-SB simulation [Matsuyama et al. 1999] for 1987-88 from GSWP (Global Soil Wetness Project; Dirmeyer [1997]). 10-day averages are resampled for monthly averages (Figure A.2).

Isotopic balance equations are described by the similar form as equation (A.1),

\[
\delta s_{(t+\Delta t)}S_{(t+\Delta t)} = \\
\delta s_{(t)}S_{(t)} + \delta pP - \delta e_sE_s - \delta e_tE_t - \delta e_iE_i - \delta rR \quad (A.5)
\]

Only isotopically fractionated component, $\delta e_s$, is separately diagnosed as

\[
\delta e_s = \frac{\delta s^* - (f^{\alpha-1}(1 + 10^{-3}\delta s^*) - 1) \times 10^3 \times f}{1 - f} \quad (A.6)
\]

where $S^*$ and $\delta s^*$ denote storage and its isotopic ratio just before evaporation from soil surface occurs, $\alpha = 0.9907$ for liquid to gas at 25°C, and $f = (S^* - E_s)/S^*$. 

Figure A.1: Schematic representation of Iso-LSM and Iso-TRIP
A. ISOTOPE RIVER MODEL

Figure A.2: 10-day averages of weighting rates of evaporation from soil ($E_s$), transpiration from vegetation ($E_t$), and evaporation from canopy interception ($E_i$). Data are reprocessed from JMA-SiB results on GSWP1.

A.2 Development of an Isotopic River Routing Scheme (Isotrip)

An isotopic module is added to total runoff Integrated pathways (TRIP) developed by Oki and Sud [1998]. The isotopic module is described as follows:

$$\delta s_{r(t+\Delta t)}S_{R(t+\Delta t)} = \delta s_{r(t)}S_{R} + \sum_{n=1}^{8} (\delta s_{r,n}O_{n}\Delta t) - \delta s_{r}O\Delta t + \delta r R \quad (A.7)$$

where $S_R$, $O$, and $R$ are water storage in river path, outflow rate from a grid, and runoff from equation (A.1), respectively; $\delta s_r$ denotes isotopic ratio of $S_R$; $\delta s_{r,n}$, $O_n$ indicate isotopic ratio of river water and

outflow rate on each adjacent grid (8 directions), respectively. A schematic is shown in the right figure in Figure A.1 and the numerical river route is shown in Figure A.3.

A.3 Results

Figure A.4 shows time scale variations in observed $\delta^{18}O$ of Ping river water and precipitation, and precipitation amount in Chiangmai, Thailand in 2001. Isotopic composition in river water relatively steadies than that in precipitation, but there is certain seasonality and a clear signal in August, when isotopically depleted precipitation took place.

Figure A.5 displays temporal variations in simulated precipitation $P$, evapotranspiration $E$, water storage $S$, and runoff $R$ at a grid includes Chiangmai. Both evaporation and precipitation increases when rainy season in Thailand starts on May. Eventually water storage increases and reaches 150 mm. Then, some of water flows as runoff since mid of May until the end of October.

Figure A.6 shows variations in respective isotopic compositions of $P$, $E$, $S$, and $R$. Fluctuation of evapotranspiration isotopes is not so large as that of precipitation.
Figure A.3: Map of Indochina Peninsula with numerical river flow routes of Chaophraya River

Figure A.4: Observed isotopic compositions of Ping river water and precipitation, and precipitation, Chiangmai in 2001

Finally, in Figure A.7, simulated and observed isotopic compositions of Ping river water in Chiangmai are compared. Fully flat line until May is resulted from the no-runoff situation during this period. This unrealistic runoff is inherently caused by a simple bucket model, and isotopic results during this period have not been used for a comparison with the observations.

Correlation coefficient is 0.62, but there is a consistent discrepancy. The simulated results underestimates by about 5 \text{‰}. However the discrepancy, the isotopic depletion on the mid of August is well reproduced by the simulation. It implies the model well simulates isotopic variation in river water for the first order estimate.
Figure A.5: Simulated precipitation, evapotranspiration, water storage, and runoff, Chiangmai in 2001

Figure A.6: Simulated isotopic compositions of precipitation, evapotranspiration, water storage, and runoff, Chiangmai in 2001

Figure A.7: Simulated isotopic compositions of Ping river water and water on the estuary of Chao Phraya
Appendix B

Review on Precipitation Recycling—Comparison of CMA results with other analytical methods

Precipitation recycling has been studied by mainly hydrologist who tries to understand atmospheric processes in addition to traditional surface processes. Colored moisture analysis (CMA) in this paper can be regarded one of the models, so that reviews on and comparison with previous studies will be described in this section.

B.1 Recycling Models

B.1.1 Budyko’s [1974] basic concept

Budyko [1974] defined that the term recycle is precipitation that originated from the specific local regions as evaporation, and that recycling ratio, $\zeta$, is the averaged local-originated precipitation $\langle P_l \rangle$ divided by averaged total precipitation $\langle \bar{P} \rangle$. Given that “well-mixed atmosphere” is assumed,

$$\zeta \equiv \frac{\langle P_l \rangle}{\langle \bar{P} \rangle} = \frac{\bar{W}_i}{\bar{W}}. \quad (B.1)$$

As in equation (2.4), the atmospheric budget equations is described as

$$\frac{\partial W}{\partial t} = -\nabla \cdot \bar{Q} + E - P \quad (2.4)$$

where $W$, $\bar{Q}$, $E$, and $P$ are vapor amount, vapor flux vector, evaporation, and precipitation, respectively. If local and remote vapor can be distinguished as $W_l$ and $W_r$,

$$\frac{\partial W_l}{\partial t} = -\nabla \cdot \bar{Q}_l + E_l - P_l \quad (B.2)$$

$$\frac{\partial W_r}{\partial t} = -\nabla \cdot \bar{Q}_r - P_r \quad (B.3)$$

Given constant wind speed, $u$ from $x$-direction, constant precipitation and evaporation in time and space, and steady condition for local and remote vapor (no change in time, $\partial W_l/\partial t = \partial W_r/\partial t = 0$).
\( \partial W_r / \partial t = 0 \), following equations are introduced:

\[
\frac{duW_i}{dx} = E - P_i
\]

therefore, \( uW_i = (E - P_i)x \) \hspace{1cm} (B.4)

\[
\frac{duW_r}{dx} = -P_r
\]

therefore, \( uW_r = uW_o - P_r x \) \hspace{1cm} (B.5)

where \( W_o \) is the vapor amount at \( x = 0 \). Therefore averaged vapor contents for the horizontal one-dimensional region with length \( L \) are expressed as

\[
W_i = \frac{(E - P_i)L}{2u} \hspace{1cm} (B.6)
\]

\[
W_r = W_o - \frac{P_r L}{2u} \hspace{1cm} (B.7)
\]

The recycling ratio \( \zeta \) is defined as

\[
\zeta = \frac{W_i}{W_i + W_r} = \frac{(E - P_i)L/2}{uW_o + (E - P)L/2} \hspace{1cm} (B.8)
\]

Substituting \( P = P_i/\zeta \), and finally the equation is solved for \( \zeta \) to give

\[
\zeta = \frac{EL}{2uW_o + EL} = \left( 1 + \frac{2uW_o}{EL} \right)^{-1} \hspace{1cm} (B.9)
\]

\( uW_o \) is the incoming moisture flux, often termed as \( F^+ \).

**B.1.2 Brubaker et al.’s [1993] method**

*Brubaker et al.* [1993] extended Budyko’s model to a two-dimensional land region, such as

\[
\zeta = \left( 1 + \frac{2F^+}{EA} \right)^{-1}. \hspace{1cm} (B.10)
\]

\( F^+ \) is defined as

\[
F^+ \equiv -\int_{\lambda_{in}} \hat{n} \cdot \hat{n} \lambda d \lambda. \hspace{1cm} (B.11)
\]

where \( \lambda_{in} \) is the sets of boundary segments and \( \hat{n} \lambda \) is the outward unit normal vector.

**B.1.3 Eltahir and Bras’s [1994] method**

*Eltahir and Bras* [1994] suggested a numeric solver to estimate the recycling ratio for region which consists of grids. Assume “well-mixed” situation on each grid, then according to the local water contents ratio (which is identical to the recycling ratio), local and remote components of moisture flux to the adjacent grid is calculated. The mathematics can be described as

\[
\zeta = \frac{P_i}{P_r + P_i} = \frac{W_i}{W_r + W_i} = \frac{O_i}{O_r + O_i} \hspace{1cm} (B.12)
\]
where $P$, $W$, and $O$ indicate precipitation, moisture content, and outflow from the grid, respectively, with suffixes indicating local ($l$) and remote ($r$).

Regarding a closed vapor budget, conservation equations for local and remote water are obtained as follows:

\begin{align}
I_l + E &= O_l + P_l \\
I_r &= O_r + P_r
\end{align}

(B.13)  

(B.14)

where $I$ indicates inflow moisture flux. These give the new recycling ratio regarding the conservation equations

$$
\zeta^* = \frac{I_r + E}{I_l + E + I_r}.
$$

(B.15)

For all grids of the region, calculation is iterated until the difference between $\zeta$ and $\zeta^*$ converges to zero.

B.1.4 Yoshimura et al.’s [2004b] method

According to the CMA results in Yoshimura et al. [2004b], recycled precipitation over total precipitation for each geographical section in Figure 6.1 is calculated.

\begin{equation}
\zeta = \frac{\int_I \int_A PdA dt}{\int_I \int_A PdA dt}
\end{equation}

(B.16)

B.2 Comparison of CMA results with other models

Figures B.1 show the 7-month averages of the recycling ratios of each region with three different methods. The recycling ratio is highly sensitive with area size of the target region. Therefore comparison of the same region with different methods has meaning. In general, Brubaker et al.’s method underestimates than the others. Also in Figure B.2, relationship of recycling ratios with corresponding area sizes are shown.

B.3 Summary

Table B.1 summarizes the characteristics of the recycling models including the one in this study.
Table B.1: Comparison of precipitation recycling related studies

<table>
<thead>
<tr>
<th>Model</th>
<th>Category</th>
<th>Strength</th>
<th>Weakness</th>
</tr>
</thead>
<tbody>
<tr>
<td>B93&lt;sup&gt;a&lt;/sup&gt;</td>
<td>Analytical</td>
<td>Few parameters required ( F^+ &amp; E )</td>
<td>Based on more than monthly averages</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Analyzable with point observation</td>
<td>In-boundary processes neglected</td>
</tr>
<tr>
<td>EB94&lt;sup&gt;b&lt;/sup&gt;</td>
<td>Analytical</td>
<td>Considering grid-based spatial variability</td>
<td>More than monthly averages</td>
</tr>
<tr>
<td>DB99&lt;sup&gt;c&lt;/sup&gt;</td>
<td>Reanalysis</td>
<td>Realistic trails of air parcel</td>
<td>Impersuasive specification for the parcel origin</td>
</tr>
<tr>
<td></td>
<td>/Lagrange</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N99&lt;sup&gt;d&lt;/sup&gt;/</td>
<td>AGCM</td>
<td>Free from “well-mix”</td>
<td>Computationally expensive</td>
</tr>
<tr>
<td>BS02&lt;sup&gt;e&lt;/sup&gt;</td>
<td></td>
<td>Detailed atmospheric processes</td>
<td>Dependent on the model bias</td>
</tr>
<tr>
<td>Y04&lt;sup&gt;f&lt;/sup&gt;</td>
<td>Reanalysis</td>
<td>Realistic moisture circulation</td>
<td>Global forcings required</td>
</tr>
<tr>
<td></td>
<td>/Euler</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<sup>a</sup>: Brubaker et al. [1993]
<sup>b</sup>: Eltahir and Bras [1994]
<sup>c</sup>: Dirmeyer and Brubaker [1999]
<sup>d</sup>: Numaguti [1999]
<sup>e</sup>: Bosilovich and Shubert [2002]
<sup>f</sup>: Yoshimura et al. [2004b]
Figure B.1: Recycling ratios of each region comparing (a) Brubaker et al.’s [1993] method, (b) Eltahir and Bras’s [1994] method, and (c) Yoshimura et al.’s [2004b] method, for April to October 1998.
Figure B.2: Recycling ratios from three different methods using the same data (GAME reanalysis) for April to October 1998, comparing with corresponding area size (logarithm).
Appendix C

Evaluation of Atmospheric Water Circulation Field in JRA-25 with Precipitation Isotopes

C.1 JRA-25

Japanese 25-year Reanalysis Project (JRA-25) has been started since 2001, and some of the data are obtainable. Also, Yoshimura et al. [2004a] shows that precipitation isotopes can be useful tools for evaluation of reanalyzed atmospheric circulation fields. This section implemented the isotope circulation model simulation with JRA-25 for certain period, and compares the results with the observations and other simulated results, i.e., NCEP/NCAR reanalysis and ERA15.

C.2 Simulation design

T106 forecasted meteorologic fields (published as fct_phy2m* files) are forced to the Rayleigh-type isotope circulation model [Yoshimura et al., 2003]. Precipitation (large scale condensation and convective precipitation), latent heat flux, vertically integrated moisture flux, and vertically integrated moisture status (precipitable water) in 6-hourly are used to simulate two streams, i.e., November 1978 to December 1983 and October 1989 to December 1994. Estimated precipitation isotopes fields are integrated to monthly values and compared with GNIP database [IAEA/WMO].

C.3 Results

Figure C.1 shows distribution of correlation coefficients between simulated and observation with more than four comparable pairs of data. Over low latitude, the temporal and spatial variations are generally well reproduced, but over middle to high latitude, the results are inferior to those of ERA15 results. (Temporal variations are shown on Figures 5.8 in Chapter 5)
The reason of the differences in good or bad agreement with the observed precipitation isotopes, Yoshimura et al. [2004a] stated the accuracy of precipitation fields. However, it is known that the precipitation in JRA25 are far better than other reanalysis products (Yamazaki, personal communication). Figures C.2 shows global distributions of correlations between daily precipitation amounts of ERA15 and NCEP/NCAR with JRA25. Theses figures are not so different each other, and do not tell us efficiently. Possible answer is that very small differences derived from characteristics of the model or assimilation schemes in each meteorologic components of the forcings become innegligible by integrating these components to simulate the large scale hydrologic circulation.

Figure C.1: As Figures 3.3. Global distribution of correlation coefficients between GNIP observations and simulated monthly results from JRA-25, for January 1979 through December 1993 and October 1989 through December 1995. Circles indicate observational sites with statistically significant correlations at a level exceeding 99%, triangles denote a level exceeding 90%, and crosses show sites where the null hypothesis cannot be rejected at the 90% confidence level.
Figure C.2: Global distribution of correlation coefficients of daily precipitation: Correlation between JRA25 and (a) NCEP/NCAR and (b) ERA15.
Appendix D

Technical notes for ICM/CMA modeling

D.1 Automatic iteration of moisture transfer calculation

Decritized form of water balance equation for isotopes (2.8) or for tagged water (6.2) should satisfy below condition every time step for stability of the numerical simulation.

\[ W - \left( \frac{|Q_x|}{\Delta x} + \frac{|Q_y|}{\Delta y} \right) \Delta t \alpha P \Delta t > 0 \]  \hspace{1cm} (D.1)

where \( W \) is total precipitable water and \( Q_x \) and \( Q_y \) are vertically integrated horizontal moisture fluxes (zonal and meridional, respectively). Precipitation \( P \) with isotopic fractionation factor \( \alpha \) should also be considered (in CMA, \( \alpha = 1 \)). If equation (D.1) is violated, new timestep \( \Delta t_2 \) is set depending on required iteration number \( I \) as,

\[ I = \text{int} \left( \left( \frac{|Q_x|}{\Delta x} + \frac{|Q_y|}{\Delta y} \right) \Delta t + \alpha P \Delta t \right) / W + 1 \] \hspace{1cm} (D.2)

\[ \Delta t_2 = \Delta t / I \] \hspace{1cm} (D.3)

In practical simulations, isotopic ratios (or tagged water contents) of incoming and outgoing moisture fluxes remain same (using the values of previous time step) during the iteration, due to avoid inconsistency between adjacent grids.

D.2 Calculation for the poles

Near the poles, decrease of \( \Delta x \) in equation (D.1) makes the CFL condition more strict. Yoshimura et al. [2003, 2004a and 2004b] did not calculate near the poles, so that it used constant values as boundary conditions. Zonal calculations for the poles (higher latitude than 85NS) are therefore introduced,

\[ \bar{W} = - \frac{\partial Q_y}{\partial y} - \bar{P} + \bar{E} \] \hspace{1cm} (D.4)

\[ \bar{R}_w \bar{W} = - \frac{\partial R_w Q_y}{\partial y} - \alpha R_w \bar{P} + \bar{R}_w \bar{E} \] \hspace{1cm} (D.5)
where bars indicate zonal averages, $R$ is the isotopic ratio or the tagged water contents ratio.

Figures D.1 and D.2 show monthly averages of precipitation $\delta^{18}O$ northward of 60N and southward of 60S, respectively, comparing between with and without consideration of the calculation for the polar regions. The northern hemisphere results on simulation with poles show systematic increase of $\delta$-values, on the other hand, slightly more depleted near the southern pole. In both figures, the differences are not spatially propagated very much, at most 70NS, so that it hardly influences lower latitude regions.

Figure D.1: Averaged precipitation $\delta^{18}O$ over high latitude regions in northern hemisphere in 1998 May (a and d), August (b and e), and April to October (c and f). (a)–(c) are the results of the new model that considers polar regions. (d)–(f) are the results of without consideration of the poles (same as the control simulation in Yoshimura et al. [2003]). Both simulations used GAME reanalysis for forcing during April and October 1998.
Isotopic fractionation from sea surface

Craig and Gordon [1965] described kinetic effect of fractionation with evaporation from water surface. ICM regards sea water as big water reservoir with constant (or climatologically variable) isotopic ratios. Evaporative isotopic ratio, $R_E$, from sea surface is written as,

$$R_E = \alpha_k \frac{\alpha^* R_s - h R_o}{1 - h}$$  \hspace{1cm} (D.6)

where $\alpha_k$ is the kinetic fractionation factor and $R_s$ is the isotopic ratio in sea surface water (usually $= 1$, same as SMOW).
Appendix E

Technical notes for Iso-MATSIRO modeling

E.1 Formation of kinetic fractionation equations and their impacts

Craig and Gordon [1965] modeled the isotopic behavior on the surface with consideration of molecular diffusivity as.

\[ \delta_E = \frac{\alpha^* \delta_w - h \delta_a - \epsilon}{1 - h + \Delta \epsilon / 1000} \]  \hspace{1cm} (E.1)

where \( \delta_E \), \( \delta_w \), and \( \delta_a \) are isotopic \( \delta \)-values for evaporation flux, reservoir, and ambient vapor, respectively, and \( h \) is relative humidity. \( \alpha^* \) is equilibrium fractionation factor between liquid and vapor (thus less than 1, usually), \( \epsilon = \epsilon^* + \Delta \epsilon \) and \( \epsilon^* = (1 - \alpha^*) \times 1000 \), and \( \Delta \epsilon \) is an additional diffusive isotopic fractionation factor, which is described as

\[ \Delta \epsilon = (1 - h) \cdot \theta \cdot n \cdot \epsilon_k \]  \hspace{1cm} (E.2)

where \( \epsilon_k \) is a kinetic constant determined by molecular number with values of 25.1\%/oo and 28.5\%/oo for \( \deltaD \) and \( \delta^{18}O \), respectively [Merlivat, 1978]. \( \theta \) and \( n \) are parameters for various surface conditions (e.g., Gat [1994], Gat [1996]).

Equation (E.1) is transformed by using following relationships:

\[ \delta = (R - 1) \times 1000 \]  \hspace{1cm} (E.3)

\[ \theta \cdot n \cdot \epsilon_k = (1 - \alpha_k) \times 1000 \]  \hspace{1cm} (E.4)

\[ R_E = \frac{1}{2 - \alpha_k} \frac{\alpha^* R_w - h R_a}{1 - h} \]  \hspace{1cm} (E.5)

Merlivat and Jouzel [1979] (MJ79) regards the kinetic fractionation term \( 1/(2 - \alpha_k) \) is a function of the wind speed, such as

\[ \frac{1}{2 - \alpha_k} \equiv \alpha_{k-MJ} \equiv 1 - \lambda \beta_k \]  \hspace{1cm} (E.6)

\[ \beta_k = \begin{cases} 0.066 & \text{for } V \leq 7 \text{ [m/s]} \\ 0.000285 \times V + 0.00082 & \text{for } V > 7 \text{ [m/s]} \end{cases} \]  \hspace{1cm} (E.7)
\[
\lambda = \begin{cases} 
1 & \text{for } ^{18}\text{O} \\
0.88 & \text{for } ^{2}\text{H} 
\end{cases} 
\]  
(E.8)

Current version of Iso-MATSIRO adopts MJ79 method, but equation (E.5) with \( \theta = 1 \) and \( n = 0.67 \) is used for some sensitivity tests (hereafter CG65 test). Results on the three sites (Manaus, Munich, and Tumbarumba) with new kinetic fractionation factors are shown in Figures E.1. Results without kinetic fractionation \((1/(2 - a_k) = 1)\) are also shown in Figures E.2. Comparing with Figures 4.5, the \( \delta^{18}\text{O} - \delta^2\text{H} \) slopes are consistently smaller than those of MWL in CG65 test, indicating stronger kinetic fractionations, but most reservoirs and fluxes are on the slope of MWL in without kinetic fractionation test.

Figure E.1: Results using different kinetic fractionation factors in Iso-MATSIRO. Seasonally averaged \( \delta^2\text{H} - \delta^{18}\text{O} \) relationship for modeled reservoirs (closed symbols) and fluxes (open symbols) at Tumbarumba (a), Manaus (b), and Munich (c). Size of each symbol shows relative magnitude of corresponding water reservoirs or fluxes. The gray line indicates the meteoric water line \((\delta^2\text{H} = 8 \times \delta^{18}\text{O} + 10)\).
Figure E.2: As Figures E.1, but results of without kinetic fractionation

E.2 Isotopic limiters for CFL condition

E.2.1 Surface isotopic fluxes and reservoirs

Isotopic fluxes of evaporation and sublimation from surface soil and canopy intercepted reservoir and transpiration from vegetation must not take all isotopes in corresponding reservoirs.

The basic surface isotopic flux is expressed in equation (4.3). Given the amount of isotopes in corresponding reservoir, $R \times W$, and the time step, $\Delta t$, following equation must be satisfied.

$$R \times W > F_{iso} \times \Delta t$$

(E.9)

Equation (4.3) is substituted to Eq. 9 to form:

$$R \times W - \frac{C_{F_{iso}}}{C_F} \left[ \frac{q^*(T)\alpha(T) R}{q^*(T) - q_a} \right] F \Delta t > \frac{C_{F_{iso}}}{C_F} \left[ \frac{q_a R_a}{q^*(T) - q_a} \right] F \Delta t$$

$$W - \frac{\alpha_k\alpha(T)}{1 - h} F \Delta t > 0$$

(E.10)

(E.11)
where $\alpha_k = C_{F,k}/C_F$ and $h = q_h/q_{(T)}$. When equation (E.11) is violated, $R_F = R$ (no fractionation) is assumed.

### E.2.2 Freeze and melt in reservoirs

Equilibrium isotopic fractionation between ice and liquid water is taken into account in soil and snow reservoirs. To avoid negative amount of isotopes in each phase, below condition is satisfied.

$$R_{liq}W_{liq} = RW - R_{ice}W_{ice} > 0 \quad (E.12)$$

where $R$ indicates isotopic ratio of reservoir $R$ with corresponding suffixes. When E.12 is violated, isotopic fractionation is not occurred ($R_{liq} = R_{ice} = R$).
References


References


References


References


