

Three dimensional modeling of Titan's aerosols and winds

by

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Three dimensional modeling of Titan's aerosols and winds

Thesis directed by Prof. Owen B. Toon

Titan's atmosphere is enshrouded by an organic aerosol haze that obscures the surface at visible wavelengths. Elucidating the nature of this haze is key to understanding Titan's complex climate system and seasonal cycles. To approach this problem, I used a global circulation model coupled to an aerosol microphysical model to explore the physical properties of the haze, its spatial and temporal distribution, and any effects on the atmosphere. I established a best-guess set of microphysical properties that describes the aerosol in Titan's atmosphere based on sensitivity tests of the parameters. From this approach I confirmed that the aerosol haze is comprised of aggregate particles with a fractal dimension of about 2. A charge on the particles equal to 7.5 electrons/micron radius best fits observations of phase function and number density, and a production rate of 10^{-14} g/cm²/s best matches vertical extinction profiles in Titan's atmosphere. I also present a formation mechanism for Titan's detached haze layer based on a balance between the vertical winds and particle fall velocities, and use a simple analytical model to reproduce the mechanism and match it to vertical extinction profiles from Cassini observations. Our simulations suggest that the detached haze layer will reappear at high altitude, around 550 km, between mid 2014 and early 2015. Finally, we show how the addition of topography and an ad hoc acceleration in our model affects the surface winds, making them more aligned with the dune crestline orientations on Titan. Through analysis of model output and comparison with spacecraft observations, I have been able to provide a coherent picture for the origin and evolution of Titan's mysterious haze.

Dedication

To my parents who have always encouraged me.

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Chapter 1

Introduction

Titan is a bizarre and fascinating world that is still struggling to be understood by modern science. It's a wild place with methane rain, haze so thick it obscures the surface, and dunes made of organic material that fell out of the sky like snowflakes. I am interested, in particular, in the aerosols in Titan's atmosphere because they affect both the climate and surface properties. Uncovering the mysteries surrounding Titan's aerosol haze provides explanatory power for many observable phenomena such as the layer of detached haze that surrounds the planet. Understanding the processes behind Titan's aerosol shroud and climate system will illuminate not only Titan, but other atmospheric planetary bodies as well. In that sense, Titan is a great laboratory in which we can test our understanding of these physical processes under conditions that we cannot replicate here on Earth.

1.1 History of Titan observations

Discovered by Christian Huygens in 1655, Titan is the largest moon in the Saturnian system. Predictions of an atmosphere came in 1925 when James Jeans calculated that it was theoretically possible for Titan to have an atmosphere if the molecules that constitute the atmosphere were heavier than hydrogen and helium [Jeans, 1925]. The atmosphere of Titan was first confirmed in 1944 by Gerald Kuiper from spectra of Titan that revealed the presence of gaseous methane [Kuiper, 1944]. Because Titan never subtends more than an arcsecond, which is similar in magnitude to the astronomical seeing due to atmospheric turbulence,

even the best ground-based telescopes were unable to resolve any features on Titan until recently. The Voyager spacecraft took close up photographs of Titan's atmospheric haze as they flew by Titan in 1980 and 1981. However, the organic aerosols in Titan's atmosphere completely obscure the planet's surface at visible wavelengths. Consequently, Titan's surface remained a mystery until the 1990s, when infrared observations from the Hubble telescope and the development of adaptive optics for ground based telescopes allowed scientists to resolve features on the Titan's surface.

In 2004, the Cassini spacecraft arrived at Saturn, and knowledge of Titan grew rapidly with new observations. Cassini carried with it the Huygens probe, which descended into Titan's atmosphere in January of 2005, providing a brief but illuminating set of in-situ data about Titan's atmosphere, as well as up-close images of its surface. So little was known about Titan's surface that Huygens was designed to float in the event it landed in a lake or ocean. Huygens sent back data during its descent and landing, with a total mission length of less than three hours. The most relevant observations to this thesis came from the Descent Imager and Spectral Radiometer (DISR), whose up- and down-looking radiometers constrained many physical properties of the organic haze enshrouding Titan.

As of 2014, Cassini is still in orbit around Saturn and making occasional close flybys of Titan. Fig. 1.1 shows two images of Titan taken by the Cassini Imaging Science Subsystems (ISS) instrument on May 7 and 8, 2012. The left image uses an infrared filter to see surface features. The thick organic haze on Titan can be seen at the limb of the planet as well. The right image shows Titan with Saturn in the background. Cassini observed lakes and seas in the polar regions on Titan, 90% of which are located in the northern polar regions. The lakes, along with observations of surface darkening associated with clouds on Titan, provide concrete evidence for the long-hypothesized methalogical cycle on Titan. Analogous to the movement of water through Earth's hydrological cycle, methane and ethane evaporate from Titan's lakes and seas, form clouds, and precipitate out onto the surface carving river valleys and making deltas. Cassini also observed large dune fields covering much of Titan's tropics.

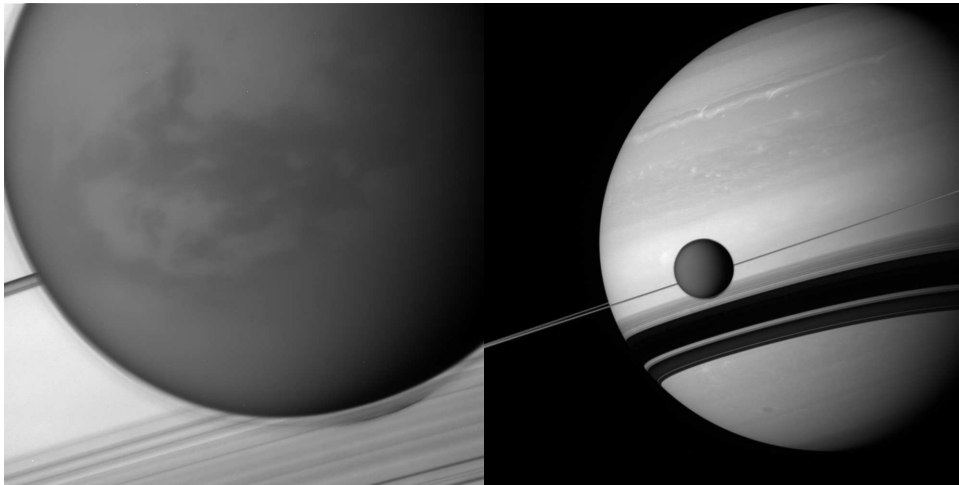


Figure 1.1: Images of Titan with Saturn in the background taken by the Cassini ISS instrument on May 7th and 8th, 2012. The left image is taken using the infrared filter CB3, which exploits a gap in the methane absorption spectrum, to observe surface features. The right image is taken with the red filter.

The continuous monitoring of Titan from Cassini over the past decade has furthered our knowledge of the seasonal cycles that are at work on Titan.

1.2 Titan properties

Titan's physical and orbital properties provide the basis for understanding its atmosphere. By comparing with Earth, we can build intuition and better understand how these properties affect Titan's atmosphere. Titan, in many ways, such as its surface features and atmospheric composition, is one of the most Earth-like bodies in the solar system. However, in many other regards, such as its bulk composition, surface temperature, and orbital properties, Titan is vastly different from Earth. Table 1.1 provides a list of Titan's physical and orbital properties side-by-side with those of Earth, along with their effect on the atmosphere of Titan.

The physical properties that affect Titan's atmosphere include its mass, radius, and composition. With a radius of 2575 km, Titan is larger than the planet Mercury. Despite being the second largest moon in the solar system, a mass of only 1.3×10^{23} kg leads to a surface gravity that is only 1/7th that on Earth. This results in an extended atmosphere with a much larger scale height than on Earth. The surface pressure on Titan is about 1.5 bars, or 50% greater than Earth, which is much closer than Venus with 90 bars and Mars with 0.007 bars. However, due to Titan's lower gravity, the column mass of atmosphere is 10 times higher on Titan, making the atmosphere sluggish. Finally, Titan has a nitrogen-based atmosphere with a small but important contribution from methane. The methane on Titan is photodissociated at the top of the atmosphere resulting in vigorous photochemistry that leads to a host of organic molecules as well as solid organic aerosol particles. An aerosol is a suspension of particles in a gas.

Table 1.1: Comparison of physical and orbital properties of Earth and Titan.

Property	Titan	Earth	Effect on Titan's atmosphere
Mass (kg)	1.3×10^{23}	6×10^{24}	Low gravity
Radius (km)	2575	6371	Low gravity
Atm. column mass (kg/m ²)	10^5	10^4	High pressure
Surface gravity (m/s ²)	1.35	9.8	Large scale height
Surface pressure (bar)	1.5	1	Long response time
Atm. composition	97% N ₂ , 2-5% CH ₄	77% N ₂ , 22% O ₂	Organic haze production
Semi-major axis (AU)	9.5	1	Low temperature
Eccentricity	0.0565	0.0167	Asymmetrical seasons
Orbital period (yrs)	29.5	1	Long timescales
Rotation period (days)	16	1	Cyclostrophic balance
Obliquity	26.7	23.3	Seasons

With an eccentricity of 0.0565, Saturn is in an elliptical orbit around the Sun at a distance of 9 to 10 AU, causing an asymmetry in the seasons. At a distance of almost

10 AU, Titan receives 1/100th of the solar insolation that the Earth receives leading to surface temperatures near 93 K. Another effect of this distance is that Saturn and Titan take 29.5 years to orbit the sun. This is a long timescale, which makes observing Titan's seasonal cycles difficult. One year on Titan is almost an entire career length for the scientists studying it. Titan's rotation period is 16 days and it is in synchronous orbit, meaning that the same side of Titan always faces Saturn. Similarly, our Moon is in synchronous orbit around Earth. Titan's long rotation period has a large impact on its atmospheric dynamics. Since Titan rotates slowly, the Coriolis force is weak and it is the inertial forces, from the zonal wind moving around the planet, that balance the pressure gradients due to equator to pole temperature gradients. On Earth rotational forces balance pressure gradients. The balance between inertial and pressure forces on Titan is known as cyclostrophic balance. Titan has an obliquity of 26.8 degrees, which is very similar to Earth's obliquity. Like all other planetary bodies with a significant axial tilt, Titan's obliquity causes seasons. Periapsis, Titan's closest approach to the sun, occurs just after the summer solstice, when the southern hemisphere is tilted toward the sun. Titan's southern hemispheric summers are shorter and more intense than the longer, more mild, northern hemispheric summers. This discussion of physical and orbital properties and their comparison with Earth helps build the framework for our understanding of Titan's atmosphere.

1.3 Motivation for atmospheric modeling

The driving goal of this work is to understand the physical nature of Titan's organic haze and its interaction with the atmosphere and surface. We need to understand Titan's aerosol to understand its climate as a whole. The aerosol particles on Titan are so abundant that they obscure the surface at visible wavelengths. They also have massive effects on the atmospheric and surface temperature due to the absorption and radiation of light. Absorption of solar UV and visible light by the haze causes upper atmosphere warming, similar to the role of ozone on Earth. This warming creates a stratosphere. Not only does the aerosol

heat the stratosphere, but it also has an anti-greenhouse effect that cools the surface by several Kelvin. Titan's haze is dynamic and heterogeneous. There are winter polar hoods, detached layers, and thick deposits at the surface. The reflectivity of the haze changes seasonally as Titan orbits the sun. One of the earliest seasonal changes observed was an albedo dichotomy between the hemisphere. One hemisphere on Titan is brighter than the other, and the bright hemisphere switches with the seasons. Figure 1.2 shows two images of Titan taken by the Cassini ISS instrument. The left image is a true color composite using the red, green, and blue filters. The planet looks yellowish due to efficient absorption at short wavelengths by the haze. On the right is a backlit image of Titan showing the high altitude detached haze as a ring around the limb of the planet. The polar hood where the detached haze layer merges with the main haze layer can be seen at the top of the image on the right.

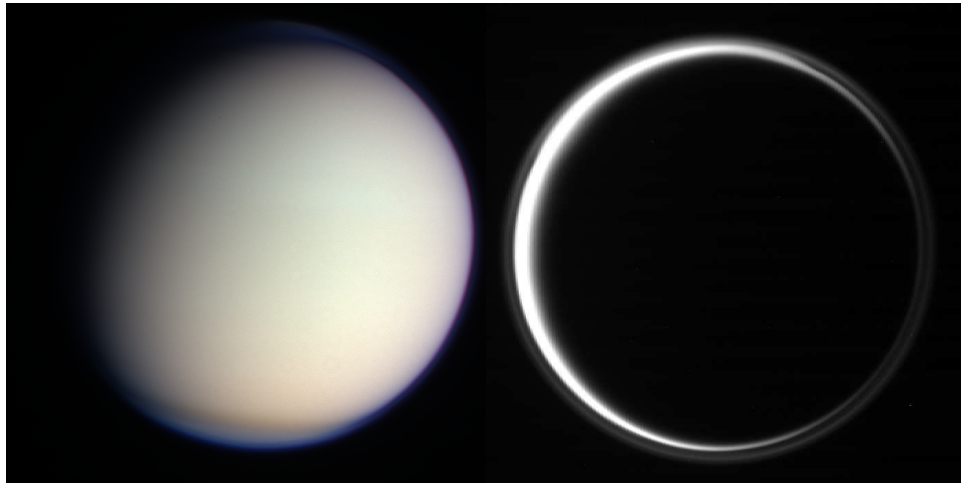


Figure 1.2: Images of Titan taken by the Cassini ISS instrument on April 9, 2012 and August 22, 2009. The left image is a true color image using the red, green, and blue filters. Notice the haze is yellowish in color. Titan's limb is illuminated in the right image making the detached haze layer clearly visible.

There have been previous efforts to model Titan's haze. Most have been one- [Toon et

al. 1992, Lavvas et al. 2010] or two-dimensional [Rannou et al. 2002, Rannou et al. 2004, Lebonnois et al. 2009]. One dimensional models do not capture the atmospheric transport, meridional variation or seasonal cycles of Titan’s haze, and while two-dimensional models are better, they still parameterize winds and fail to take into account such surface features as albedo and topography. As a result, three-dimensional global circulation models (GCMs) are needed to fully explore the processes driving Titan’s haze and resulting climatic effects. To date there is one other 3D GCM with aerosol microphysics [Lebonnois et al., 2012b], however they have not analyzed the haze properties. To address this need, I have expanded upon the National Center for Atmospheric Research (NCAR) Community Atmospheres Model (CAM) version 3. This is a climate model that was developed initially for Earth and ported to Titan by Friedson et al. [2009]. I have modeled how Titan’s organic aerosol interacts with the atmospheric dynamics and radiation.

1.4 Dissertation description

This thesis is divided into 5 chapters. The first and last chapters are the introduction and conclusion. The second chapter of this thesis introduces the Titan CAM model and new developments we have made. I coupled the microphysical model CARMA to CAM, including the radiation package. I also updated the aerosols in CARMA to include fractal physics. After much effort trying to reproduce superrotation in the model, we eventually artificially forced the dynamics to be consistent with observations. With realistic, albeit forced, dynamics we were able to explore the parameter space of the microphysical model. We identify a best fit case of aerosol properties that most closely match the suite of observations from the Huygens probe, Cassini and Voyager spacecraft, and ground based telescopes. We also discuss the effects these aerosols have on the heating, cooling, and dynamics of the atmosphere. In the third chapter we use our model to investigate the detached haze layer in Titan’s atmosphere. We attempt to understand the origin of the detached haze layer and distinguish between two competing theories of its formation. In the fourth chapter of this

dissertation we investigate the surface dune-forming winds on Titan using our GCM. We are interested in how these surface winds can explain the abundance and direction of dunes observed at low latitudes. We find that topography plays an important role in controlling surface wind direction in our model.

Chapter 2

The Titan CAM

2.1 Introduction

Organic aerosols are produced in Titan's upper atmosphere through complex photochemistry involving the destruction of methane [Yung et al., 1984, Wilson and Atreya, 2004]. The methane (2%) and nitrogen (98%) in Titan's upper atmosphere dissociate because of UV photons or high energy particles to create a variety of organic compounds including: ethane (C_2H_6), ethene (C_2H_4), acetylene (C_2H_2), hydrogen cyanide (HCN) and a host of other more complex organics. Chemical and aerosol modeling suggests that small, $0.05\text{ }\mu\text{m}$, spherical monomers are produced from the photochemistry and ensuing coagulation hundreds of kilometers above Titan's surface [Wilson and Atreya, 2004, Tomasko et al., 2008b, Lavvas et al., 2010]. As the monomers drift downward they coagulate further into fractal aggregates reaching sizes near a micron around the tropopause at 50 km above Titan's surface [Cabane et al., 1993, Lavvas et al., 2010]. As the particles coagulate and fall they likely act as seed nuclei for the condensation of methane, ethane and other condensable hydrocarbons. The aerosol particles are probably insoluble in hydrocarbons and have low volatility, so they do not evaporate [McKay et al., 2001, Coll et al., 1999]. Models suggest that only about 1% of the aerosols are removed from the atmosphere through precipitation [Barth and Toon, 2006]. The vast majority of the aerosol particles may be deposited on the surface through dry deposition.

There have been a significant number of laboratory syntheses of Titan-like organic

aerosols, which are often referred to as tholins. Tholin, a term coined by Carl Sagan and Bishun Khare, is Greek for unclear or muddy, and refers to the organic residue produced in these experiments. Most of these experiments have put methane and nitrogen into a low pressure chamber irradiated with ultraviolet light (UV) or with an electrical discharge. Khare et al. [1984] determined the real and imaginary indices of refraction of one type of synthetic tholin. These refractive indices almost match those needed to reproduce the albedo of Titan. However, they are missing near infrared bands recently observed on Titan indicating they are not a perfect match. More recent work has produced tholins with these bands [Curtis et al., 2008, Imanaka et al., 2004]. Electron microscopy of synthetic tholins found spherical monomers and fractal aggregates [Coll et al., 2001, Trainer et al., 2006].

There is a long history of observations on Titan that can be used to validate and constrain aerosol models. Rages and Pollack [1983] used Voyager data to measure the extinction profile in the upper atmosphere of Titan. They discovered the detached haze layer above 350 km. West et al. [2011] found that the height of the detached haze layer falls from over 500km to 350km around equinox. Ground based studies of Titan’s spectrum measured a steep wavelength dependence of the aerosol optical depth [Griffith et al., 1991, Gibbard et al., 1999]. Tomasko et al. [2005, 2008b] use a detailed analysis of Huygens observational data to constrain the aerosol vertical structure and wavelength dependence of aerosol optical properties including single scattering albedo and optical depth. They found that the optical depth falls off sharply with increasing wavelength, however the slope of the wavelength dependence softens below 80km. They also measured the phase functions of the aerosols, which can be used to constrain the particle size. Rannou et al. [2010] used Cassini VIMS data to produce latitudinal gradients of the aerosol optical depth in the infrared (IR).

Several processes control Titan’s geometric albedo [McKay et al., 2001]. Rayleigh scattering by gasses affects the UV. The aerosols absorb and scatter UV and optical wavelengths of light. In the IR, where the wavelength of light is larger than most of the particles, the aerosols are slightly absorbing. There are also methane bands and surface reflectance in

the near IR. Thus, the geometric albedo is controlled by Rayleigh scattering, the aerosol properties, surface properties, and the methane abundance. In the 1970's and 1980's temporal variations in Titan's albedo were measured [Lockwood, 1977, Lockwood et al., 1986]. These were first thought to be due to the solar cycle. However, Sromovsky et al. [1981] observed hemispheric asymmetry in Titan's albedo from Voyager. Different views of the planet over its orbit, as well as seasonal changes in the albedos of the two hemispheres lead to the seasonal changes in albedo. Lorenz et al. [1997, 1999] used HST to measure hemispheric asymmetry and wavelength variation in Titan's albedo. These significant latitudinal, seasonal, and wavelength dependent variations in the geometric albedo of Titan's haze, suggest complex interactions between the aerosols and dynamics. The current hypothesis is that these variations are caused by interactions between aerosol transport by settling and by winds [Toon et al., 1992, Hutzell et al., 1996, Rannou et al., 2002]. Essentially, particle sizes are larger when and where upward motions suspend the particles against falling, and smaller when and where downward winds shorten the aerosol lifetime. Tokano et al. [1999] used a three-dimensional model that lacked radiative coupling between aerosol and dynamics to demonstrate the dynamical effect on the aerosols. Rannou et al. [2004] first coupled the 2-D dynamics and microphysics showing that aerosols have a strong influence on dynamics, which can explain the detached haze layer.

The radiation absorbed, scattered, and emitted by the aerosols redistributes heat in the atmosphere creating thermal gradients. The atmosphere responds to these thermal gradients with winds that act to bring the atmosphere back to thermal equilibrium. The aerosols in the upper atmosphere reflect about 30% and absorb about 40% of the incident solar radiation making it the dominant force in controlling stratospheric temperatures and thermal gradients [McKay et al., 2001]. Due to long radiative response times relative to a Titan season in the troposphere and short response times in the stratosphere, we would expect the troposphere to have latitudinal symmetric temperature gradients while the stratosphere should show a seasonal response [McKay et al., 2001]. Tomasko et al. [2008a] used Huygens

data to produce vertical heating and cooling profiles on Titan. They found that the radiative heating rate exceeds the cooling rate at that site by 0.5 K/day. Since no evidence for changing temperatures was found, the excess heat must be transported away from the equator. Vinatier et al. [2007a], Teanby et al. [2008] used CIRS data from Cassini to retrieve vertical temperature profiles within the atmosphere. They find differences in stratospheric temperature between the hemispheres, sometimes as large as 30K. However, the warmer hemisphere switches around 250 km in altitude. Teanby et al. [2012] also used trace gas abundances in Titan's stratosphere to display the Hadley cell circulation.

Super-rotating winds have been inferred from observations by Voyager and Cassini on Titan. The velocity and distribution of the winds were retrieved via the gradient wind equation from temperature and pressure profiles from Voyager, and CIRS on Cassini, as well as Doppler shifts and solar zenith angle variation of the Huygen's descent probe [Flasar et al., 1981, Bird et al., 2005, Allison et al., 2004, Achterberg et al., 2008]. These zonal prograde winds reach speeds between 100 and 200 m/s and peak in the stratosphere at altitudes between 150 and 300 km. The upper air winds are in cyclostrophic balance and not geostrophic balance as on Earth due to the slow rotation rate of Titan [Gierasch, 1975]. There is evidence for a seasonal cycle in the super rotating winds as well [Kostiuk et al., 2010].

Several groups have used GCMs to model Titan's atmosphere. Hourdin et al. [1995] were the first to reproduce the superrotating prograde winds. They also showed that the meridional circulation is dominated by pole to pole Hadley cells in the stratosphere, except around the equinox when it becomes an equator to pole cell. Tokano et al. [1999] were able to reproduce the temperature gradients and superrotation with a model that included uniform haze opacity, but had to artificially damp the meridional circulation. Rannou et al. [2004] compared a uniform haze layer to a haze coupled to the radiation and dynamics. In their two dimensional model, they found the aerosols accumulating at the poles. The aerosols radiated in the infrared enhancing the cooling at the poles and increasing the temperature

gradient. The increased temperature gradient intensified the zonal winds. However, Rannou et al. [2004] used an aerosol production rate of $1.2 \times 10^{-13} \text{ g cm}^{-2} \text{ s}^{-1}$, which is about an order of magnitude higher than the consensus from 1-dimensional models [McKay et al., 2001]. Richardson et al. [2007] developed a planetary model based on the Weather and Research Forecasting (WRF) model. Recently, Newman et al. [2011] were able to reproduce Titan’s superrotation by eliminating all imposed horizontal diffusion in the Titan WRF. They identified the angular momentum transfer from the surface into the upper atmosphere as driving the superrotation. Although this model includes aerosols in the radiative transfer, it is a simplified scheme. Liu et al. [2008] used an older version of the Community Atmosphere’s Model, CAM2, to explore Titan’s dynamics. They were also able to reproduce the equatorial super rotation and vertical structure by forcing a latitudinal temperature gradient. They imposed, rather than solved for, a meridional thermal structure, thus they introduced the conditions to generate superrotation artificially. Friedson et al. [2009] produced a three-dimensional model for Titan using the NCAR CAM3 model to which we have coupled an aerosol microphysics model.

Toon et al. [1992] described a Titan version of the Community Aerosol and Radiation Model for Atmospheres (CARMA), a cloud microphysics and aerosol model [Toon et al., 1988]. This model solves the continuity equations for different sized aerosols at each time step and grid point taking into consideration coagulation, vertical fall velocities and sedimentation as well as other processes such as condensation growth, which are not relevant to tholins. Barth and Toon [2004, 2006] used one-dimensional versions of this code to investigate ethane and methane condensation, precipitation and cloud formation.

We combine the CAM-Titan and CARMA models to create a new model with the aerosol microphysics coupled to the dynamics and the radiative transfer. We also implemented a fractal treatment of the aerosol particles similar to Wolf and Toon [2010], which is necessary to accurately represent the particles in Titan’s atmosphere. We analyzed the three dimensional aerosol distribution and seasonal variation. Here we present the initial results

of this new model. We explore the suite of parameters that give the best fit to the aerosol data including, the vertical profile and wavelength dependence of the aerosol optical depth, number density and particle size. We also investigate how the aerosols affect the heating rates, temperature profiles, and albedo on Titan. Finally, we consider the dynamical effects of changing aerosol parameters and model resolution in the vertical and horizontal.

2.2 Model Description

2.2.1 CAM Model Description

Friedson et al. [2009] adapted the Community Atmospheres Model (CAM3) to Titan. This new model includes a dynamical core, a treatment of the planetary boundary layer (PBL), tidal forces from Saturn, radiation, surface interactions, a photochemistry package, and a microphysical treatment of the aerosols.

The Titan CAM [Friedson et al., 2009] has a dynamical core that uses a finite volume scheme that does sequential parallelization in 2 directions, latitude x longitude and then remaps fields in altitude. The model has 61 vertical levels going from the surface to 0.35 Pa or about 580 km. Most simulations were run at a resolution of 10x15 degrees in latitude and longitude. This is a coarse grid, but given the large amount of computing power required to simulate Titan’s long timescales, it is an appropriate choice. We generally have to perform runs that are several hundred Earth years in length and, due to the small size of Titan and the high wind velocities, short time steps are required to maintain stability. Typically, we run the models for 500 Earth years, or about 17 Titan years. One drawback of the coarse resolution is that the finite volume dynamical core becomes highly dissipative. This may be the primary reason that the model of Friedson et al. [2009] did not produce strong superrotation. To test this theory, we also run a simulation at 4x5 degree resolution. Titan CAM uses the method of Tokano and Neubauer [2002] to add tidal accelerations from Saturn to the winds. There is a chemistry and trace species package; however, it is not integrated

with the radiation. Calculation of the planetary boundary layer properties uses the terrestrial routine from Collins et al. [2004] with non-dimensional constants that should be universal. Calculations of sensible heat and momentum fluxes from the surface use the terrestrial code also described in Collins et al. [2004]. The surface albedo map used in the radiation code is specific to Titan. The albedo ranges from 0.14 to 0.3 from the UV to near IR, and is interpolated from the ISS map [Friedson et al., 2009].

The radiation code in the Titan CAM is broken up into two regimes, shortwave (< 3 μm) and longwave (> 5 μm). The shortwave radiative transfer code uses a two-stream delta-Eddington approximation over 20 wavelength intervals with four correlated k coefficients per interval. The shortwave opacity includes the aerosols, CH_4 , and Rayleigh scattering. The methane abundance is horizontally uniform and matches the DISR vertical profile [Niemann et al., 2005]. No chemistry is considered in these simulations. The longwave radiative fluxes are calculated using a two-stream direct integration method using 159 intervals from 10-1600 cm^{-1} with 20 correlated k coefficients each. The longwave radiatively active gasses include C_2H_2 , C_2H_4 , C_2H_6 , and HCN . We also include collision-induced absorption from $\text{N}_2\text{-N}_2$, $\text{N}_2\text{-H}_2$, $\text{N}_2\text{-CH}_4$, and $\text{CH}_4\text{-CH}_4$ as described by Friedson et al. [2009]. All of these gasses are horizontally uniform. The model uses a simple upper boundary condition for the radiation code in which the atmosphere above the top layer of the model matches the top layer in opacity.

The aerosols in Friedson et al. [2009] have a horizontally uniform prescribed abundance. The vertical abundance was used as a free parameter to match the DISR short wave heating and long wave cooling rates found by Tomasko et al. [2008a]. The earlier model used prescribed aerosol optical constants; specifically the extinction coefficient, single scattering albedo, and asymmetry parameter derived from Cassini/Huygens observations [Tomasko et al., 2008b]. We replaced horizontally uniform prescribed aerosol abundances with our CARMA calculated abundances. We replaced the older optical properties, which were created to match DISR heating results, with the optical properties for fractals calculated from

the mean field approximation theory described in Botet et al. [1997]. This makes the aerosol heating and cooling in each grid cell and level of the atmosphere a function of the particle size and number density. This functionality has important effects on the meridional temperature gradients.

The previous version of this model was unable to reproduce Titan’s superrotating winds [Friedson et al., 2009]. Several attempts were made to increase the zonal wind speed through a variety of modifications listed in Table 2.1. We removed all of the imposed horizontal and vertical diffusion, as suggested by Newman et al. [2011]. We included realistic and exaggerated aerosol heating and cooling as suggested by Rannou et al. [2004]. We included topography into our GCM to overcome errors in the surface angular momentum budget. We tested higher resolutions to reduce numerical diffusion. We also reduced and removed surface drag and tested variations to the gravity wave parameterization. Other than higher resolutions which had low to moderate improvements to zonal wind speed, these changes had little effect. Our model produced zonal winds from 5-40 m/s, well below the 200 m/s inferred on Titan.

Table 2.1: Attempts to increase the zonal wind speed in the model.

Possible cause of lack of superrotation	Attempted improvement
Improper heating/cooling	Add realistic aerosol heating and cooling via CARMA
Gravity wave parameterization	Adjust or remove
Sponge layer	Reduce diffusion
Numerical diffusion	Increase resolution
Surface stresses	Reduce surface drag
Topography	Include realistic topography map from Lorenz et al. [2013]
Dynamical core	Try other cores (unsuccessful)

Lebonnois et al. [2012b] compared the angular momentum terms of several GCMs on a modified Earth, Venus, and Titan. They found that the CAM finite volume dynamical core

has errors in its angular momentum conservation that affect its ability to reproduce superrotation. However, for a simplified Earth and Venus simulations, the addition of topography increased the magnitude in the angular momentum terms at the surface, overcoming the errors and allowing superrotation to form. We tested this on Titan by adding topography based on Cassini radar measurements from Lorenz et al. [2013]. However, Titan has very small topographic changes and a highly extended atmosphere. The addition of this topography did not increase our zonal wind speed. It is our conclusion that the CAM 3 model running the finite volume dynamical core in its current form is incapable of reproducing Titan’s superrotating winds. Our work supports the conclusions of Lebonnois et al. [2012b].

Consequently, we have opted to artificially force Titan’s zonal winds into superrotation by adding ad hoc acceleration. It will be shown in section 3.8 and 3.9, that by forcing our model to have superrotating winds, we naturally establish a latitudinal temperature gradient of 20 K in the upper stratosphere. This is the opposite approach of many other models, which have forced the temperature gradients to produce the winds [Liu et al., 2008]. Titan’s latitudinal temperature gradient is expected to be related to the wind speed via the thermal wind equation [Achterberg et al., 2010]. This observational link is supported by Liu et al. [2008], who modeled Titan with a simplified version of CAM 2. Their model produced 100+ m/s zonal winds by relaxing the temperatures to prescribed values with very large (30K at 140 km) equator to pole temperature gradients. Our model is unable to reproduce large temperature gradients without this ad hoc acceleration.

We updated our current model to include an ad hoc zonal wind acceleration used to artificially increase the wind speed to match observations. This forced, but realistic, dynamics gives a better environment in which to analyze the aerosol microphysical properties. This acceleration is highest at the equator and falls off exponentially with the square of the latitude. The zonal acceleration is constant in the upper atmosphere at pressures smaller than 5 hPa. We include no zonal acceleration between 5 hPa and 100 hPa, consistent with the lack of superrotation between these levels in Titan’s atmosphere. At 100 hPa we introduce an

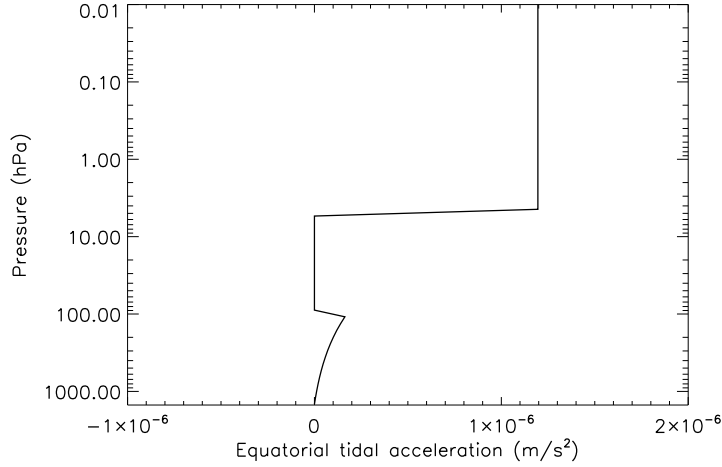


Figure 2.1: Vertical profile of the ad hoc acceleration used in the model.

acceleration that decreases as you approach the surface as seen in Fig. 2.1. The acceleration term below 100 hPa has the following form.

$$A = A_0 \exp^{-(\phi/\phi_0)^2} (100(P/P_0)^{-0.577} - 3.771) \quad (2.1)$$

Where A is the ad hoc acceleration term, A_0 is a constant with a magnitude of 1.25×10^{-8} determined from forcing the modeled winds to match observations, ϕ is latitude in radians, ϕ_0 is a reference latitude of 0.524 radians or 30 degrees, P is pressure, and P_0 is the reference pressure of 5 hPa. Above 5 hPa we use the same equation, without the pressure term. The peak magnitude of this acceleration, which occurs at the equator at pressures less than 5 hPa, is $1.2 \times 10^{-6} \text{ m/s}^2$. This acceleration is a modest fraction of the maximum tidal amplitude which reaches a value of $7 \times 10^{-6} \text{ m/s}^2$. However, the tidal accelerations cancel out over a Titan day while this ad hoc acceleration is constant in time. The force required to accelerate a column of atmosphere at the equator as we have in the ad hoc scheme described above is $4.5 \times 10^{-3} \text{ N/m}^2$. For comparison, this is within 20% of the force of the surface drag for

1m/s surface winds. The surface winds described here are considered to be at an altitude of 175 m, the middle of our lowest model layer. Our model with the ad hoc forcing gives a seasonally oscillating zonal jet peaking at 140 m/s at 0.05 hPa. The peak of our zonal jet oscillates in strength as well as latitude seasonally due to solar forcing. Results comparing the temperatures and dynamics of our model with and without the zonal wind acceleration will be shown and discussed further in section 3.8 and 3.9. Unless otherwise noted, results will be shown from microphysical simulations with the ad hoc acceleration.

2.2.2 CARMA Description

CARMA solves the aerosol continuity equation in one spatial dimension, as described by Toon et al. [1992].

$$\begin{aligned} \frac{dn(v)}{dt} = & \frac{dn(v)V_{fall}}{dz} + 1/2 \int_0^v K(v', v-v')n(v')n(v-v')dv' \\ & - \int_0^\infty K(v', v)n(v)n(v')dv' + P(v) - Rn(v) \end{aligned} \quad (2.2)$$

Where $n(v)$ is the number of aerosol particles of volume v , V_{fall} is the fall velocity, K is the coagulation kernel, P is aerosol production rate, and R is the rainout rate. The aerosols are produced high in the atmosphere, coagulate and fall, and are removed at the surface through deposition. Below and in the appendix we describe these processes in detail. Appendix A.1 provides the radius grid used in the model.

2.2.2.1 Aerosol Production

The aerosol monomers are produced from a series of poorly understood photochemical and ion interactions involving methane and nitrogen [Wilson and Atreya, 2004, Lavvas et al., 2008]. McKay et al. [2001] do a thorough job of summarizing the pre-Cassini views of the production rate of Titan's aerosols from photochemical models and laboratory experiments and find the production rate to be between 0.5 and 2×10^{-14} g/cm²/s. Lavvas et al. [2008] recently used a photochemistry model which predicts a production rate of $1.27 \times$

10^{-14} g/cm²/s, assuming a monomer size of 7.25 angstroms. Wilson and Atreya [2009] used a chemical model to come up with a production rate of 2.4×10^{-14} g/cm²/s. Rannou et al. [2004] used a production rate of 12×10^{-14} g cm⁻² s⁻¹, almost an order of magnitude higher than the other estimates, to match the aerosol optical depths in their GCM. We explore a range of production rates from 0.5, 1, 3, 5 and 10×10^{-14} g/cm²/s at the top of the model in order to simulate the observed aerosol optical depths.

In this study we also compare inputting aerosols at three different sizes; 2 nm, 42 nm (monomers), and 660 nm. The 660 nm particles are assumed to be fractals with a monomer size of 50 nm. We input the aerosols at the top of the model. The assumption of a production of monomers above the model top is based on observations from Cassini, and models such as that of Rannou et al. [2004] and Lavvas et al. [2010]. The observations and models suggest that the aerosol production zone is above or near the top altitude of our model (580 km) and that large monomers form at these high altitudes [Waite et al., 2007].

2.2.2.2 Coagulation with charging

The number of aerosols in a bin of size r_i can increase if two smaller particles coagulate to create a particle of size r_i or decrease if a particle of size r_i coagulates with another particle to form a larger particle. The rate at which these processes happen is controlled by the coagulation kernel (Fig. A.1). Coagulation happens through two processes, Brownian motion, and gravitational falling with the coagulation kernel equal to the sum of these two coefficients. Coagulation is thought to be inhibited by particle charging and the coagulation kernel can be controlled by the charge to radius ratio (see appendix A.1).

2.2.2.3 Fall velocity and Sedimentation

The aerosol particles fall due to gravitational settling until they reach the surface of Titan where they are removed. In the Stokes limit where the mean free path is smaller than the particle radius, the fall velocity is calculated by equating the force of gravity on

a sphere to the air resistance of the sphere. In this limit the fall velocity is independent of pressure, but does depend on temperature. In the kinetic limit, in which the particle is smaller than the mean free path of air, the gravitational force on the particle is balanced by the momentum transfer of individual air molecules bouncing off of it. In this limit the fall velocity is inversely proportional to pressure. The details of the sedimentation algorithms are discussed in the appendix. Fall velocity is modified for fractals relative to spheres as discussed in appendix A.2. Figure 2.2 shows the fall velocities as a function of height for the set of fractal aerosols described in appendix A.2. The transition to the different regimes is apparent in their dependencies on pressure for the different sized particles. Note the kink in the fall velocities at 40 km. This kink is due to low temperatures at the tropopause. The convergence above 100 km of fall velocities for radii between 0.05 and 1 μm creating the thick line in Fig. 2.2 is caused by the fractal nature of these particles. For a fractal dimension of 2, the velocity increase expected with a doubling of mass for different sized particles is offset by the increased drag force due to a doubling of surface area. The aerosol fall velocity in the stratosphere increases as the fractal dimension increases above two.

The dry deposition flux, which occurs only between the lowest model layer and the ground, is the product of the number of particles, n , and the deposition velocity, V_d . Our dry deposition calculation employs the fall velocities as the deposition velocities and neglects the effects of turbulent transfer, Brownian diffusion, and impaction. These effects depend on wind speed, surface roughness and particle size and tend to increase the deposition flux. On the surface of Titan, the wind speeds are less than 2 m/s and the particles are around 1 μm in size. For 1 μm sized particles in an atmosphere with friction velocities of 1 m/s, we calculate an increase in the deposition velocity above the fall velocity of a few percent. The actual friction velocity is probably an order of magnitude lower than 1 m/s, the laminar flow speed, so the increase in deposition velocity above the fall velocity is negligible on Titan. Therefore we believe our simple scheme of using the fall velocity as the deposition velocity is sufficient in this model.

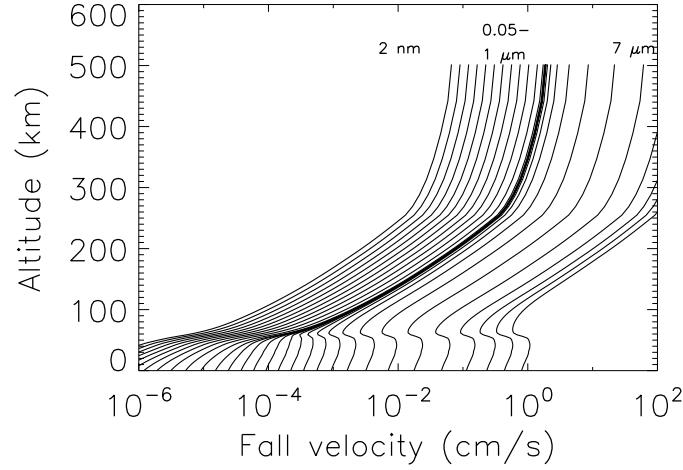


Figure 2.2: Fall velocities with height. Each line correspond to a fractal aerosol particle of a given size, as described in Table A.1 in the appendix.

Wet deposition, or rainout from methane and ethane precipitation, could be important to include in the model because the lifetime of the aerosols in the troposphere is so long. Barth and Toon [2006] found that 1% of the aerosols are removed through this mechanism, while the rest are removed by dry deposition. However they did not include fractal physics in their aerosol model. We include rainout in some model runs and explore a rainout lifetime of 50 years. It should be noted that this rainout lifetime has no physical basis because we are not simulating the ethane and methane clouds, it is just a removal mechanism of aerosols to reduce the extinction below 30 km. A similar removal could be accomplished by increasing the fractal dimension in the lower atmosphere due to condensation allowing the aerosols to fall faster. The effects of condensation will be investigated in future studies. Rainout has the effect of completely removing particles where it occurs, because it is assumed the precipitation falls rapidly to the surface. In contrast, condensing a small amount of gas and slightly increasing the particle fall velocity does not remove the particles locally. As

discussed later the shape of the observed extinction profile, is more consistent with a slight increase in the particle fall speeds in the troposphere, rather than precipitation removal.

2.2.2.4 Fractal Physics

There is abundant evidence from laboratory studies, modeling, and observations that the aerosols in Titan's atmosphere are fractal in nature. Fractal means that the aerosols exhibit the same structure on many scales, similar to fern fronds or snowflakes. The fractal aerosols are composed of monomers, small spherical aerosols with inferred sizes of 40-50 nm [Tomasko et al., 2008b]. Modeling the aerosols as fractals as opposed to spheres has several important effects. Fractal aerosols absorb more in the UV to mid-visible wavelengths than do spherical aerosols of the same mass [Botet et al., 1997, Wolf and Toon, 2010], which is important in order to match the geometric albedo of Titan. Fractal particles also have lower fall velocities than spherical particles of the same mass allowing them to stay aloft longer and coagulate into larger particles. The coagulation of fractal particles relative to equivalent mass spherical particles is enhanced by the larger radius as well.

Following the work of Cabane et al. [1993] we describe the aerosols in our model as fractal aggregates. Wolf and Toon [2010] devised the procedures for simulating fractals in CARMA. The fractal radius of the particles is given by

$$r_f = r_s^{3/D_f} r_{mon}^{1-3/D_f} \quad (2.3)$$

Here r_f is the fractal radius, r_s is the radius of an equivalent mass sphere, r_{mon} is the monomer radius and D_f is the fractal dimension. When the spherical radius is less than the monomer radius, D_f is 3, and r_f equals r_s . When r_s is greater than the monomer size, then r_f is greater than r_s . The fractal dimension is a measure of the compaction of the fractal aerosol. A fractal dimension of one would be a linear string of monomers while a fractal dimension of three would be a spherical cluster of monomers, equivalent in mass and radius

to a sphere of monomer material.

In our model, all particles smaller than the 50 nm monomer radius were assumed to be spheres, so r_f would equal r . The fractal dimension of Titan’s aerosols is not well constrained, however. Often a value of two is assumed, which leads to optical depths that produce a good match to the geometric albedo [Cabane et al., 1993, McKay et al., 2001, Tomasko et al., 2008b]. The flattening out of the wavelength dependence of the optical depth below 80 km provides evidence of a higher fractal dimension in Titan’s troposphere [Tomasko et al., 2008b]. This flattening could be due to wetting and growth of the particles as methane, ethane, acetylene or other trace gasses condense onto them. However, fractal particles are observed to compact as size increases because branches of the particles come into contact via Brownian motion. To include this effect we made the fractal dimension a function of the aerosol size, as seen in Table A.1 in the appendix. The bulk of the aerosols in our model have a fractal dimension very close to 2, with a maximum fractal dimension of 2.1 in some cases.

2.2.2.5 Radiative Coupling and Aerosol Optical Properties

CARMA currently treats aerosols as spheres or fractals. For the spheres, the optical properties are calculated using Mie theory (see appendix A.3), which solves Maxwell’s equations for spherical particles exactly. For input to the Mie code, we use the real and complex indices of refraction measured from tholins synthesized in a laboratory by Khare et al. [1984], and also the imaginary index of refraction derived from the DISR observations on the Huygens probe [Lavvas et al., 2010]. For the long wavelengths, we used imaginary indices from Vinatier et al. [2007b]. It should be noted that other models have used modified versions of the Khare indices in order to fit the geometric albedo correctly. Specifically, Toon et al. [1992] got the best fit by multiplying the imaginary index by 1.5, McKay et al. [1989] used a factor of 4/3, while Rannou et al. [1995] used a factor of 3. It seems that the Khare values of the imaginary index of refraction underestimate the actual values. The DISR data supports

this conclusion with derived refractive indices being about 20% larger at blue wavelengths and an order of magnitude larger than Khare et al. [1984] in the infrared [Lavvas et al., 2010].

To get the optical properties of the fractal aerosols we use code provided on the web by Pascal Rannou and described in Botet et al. [1997]. This code uses the mean field approximation, which approximates the Mie theory scattering coefficients a_n and b_n for each monomer inside a cluster as a common value. The electromagnetic field scattered by the aggregate is the sum of the fields scattered by each monomer. This code calculates the phase functions, scattering and absorbing coefficients, and asymmetry parameter for each particle given the monomer size, number of monomers, refractive indices of the tholin, and the wavelength in the model. The fractal particles are much more absorbing and scattering than spheres in the UV and optical wavelengths for particles with sizes near a micron as seen in Figure A.2 in the appendix.

2.3 Simulations

In this paper we present the results of several simulations exploring the sensitivity of the model to our free parameters, including the boundary conditions on haze mass production and initial particle size, aerosol shape, charge to radius ratio, and rainout lifetime in the troposphere. The parameters, their values and ranges, and comparisons with the values in a one dimensional model by Lavvas et al. [2010] can be found in Table 2.2. The base case simulation inputs $3 \times 10^{-14} \text{ g cm}^{-2} \text{ s}^{-1}$ of 42 nm aerosols which are allowed to coagulate into fractals. This production rate is a factor of three lower than the production rate Rannou et al. [2004] used in their 2-D GCM. It is consistent with the value found by Wilson and Atreya [2009] in their chemical model. At the top of our model we input spherical particles with a 42 nm radius. Our monomer size is 50 nm, so any particles larger than 50 nm will be aggregates. A monomer size of 50 nm was suggested by Tomasko et al. [2005]. However, Tomasko et al. [2009] updated the size to 40 nm +/- 10 nm. Despite this we used a 50

nm monomer radius for better comparison with other models and data sets that assume an aerosol monomer size [Lavvas et al., 2010, Tomasko et al., 2008b]. The aerosol particles in the base case simulation have a charge of 15 electrons per micron and no rainout in Titan’s troposphere. The base case aerosol parameters are the same as those used by Lavvas et al. [2010] in their 1-D model, although the 1-D model extends to higher altitude. Near 580 km, their particles are about 20 nm in radius. This choice of parameters allows easy comparison between the models as well as being a reasonable starting point to investigate the effects of the aerosols on the atmosphere as a whole. We explore these parameters using a model top of 0.35 Pa, or about 580 km. In each sensitivity test presented the only change to the model relative to the base case is the parameter we are testing. After performing our sensitivity tests, we have come up with a best guess set of aerosol parameters that are most consistent with the observations.

2.3.1 Aerosol lifetimes

As the aerosols in Titan’s atmosphere descend through the atmosphere, they can be removed in three ways. They can coagulate together to form larger aerosols, they can fall to the surface due to gravitational sedimentation, or they can be scavenged by droplets during precipitation events (rainout). We can look at the relative importance of these three mechanisms by calculating an aerosol lifetime associated with each process as shown in Figure 2.3. The rainout lifetime is fixed in this model and we explored different values to match the near surface extinction. Our best fit model has no rainout, however our base case simulation uses a rainout lifetime of 50 Earth years. The sedimentation lifetime is calculated as the time it takes the effective aerosol size (shown as the dash dot line) to fall through that vertical level in our model. The model heights are about 20 km near the top and under a kilometer near the surface. The coagulation lifetime is the time during which a monodisperse aerosol distribution would be reduced in number by a factor of 2 (calculated as two divided

Table 2.2: Parameters used in the Titan CAM model

Parameter	Base Case	Range	Lavvas et al. [2010]	Best Guess	Observations Affected
Production rate ($\text{g cm}^{-2} \text{s}^{-1}$)	3×10^{-14}	1,3,5,10 $\times 10^{-14}$	3×10^{-14}	1×10^{-14}	extinction, optical depth
Aerosol input size (nm)	42	2, 42, 663	2	42	number density in stratosphere
Top of model altitude (km)	580	380-580	1300	580	optical depth, extinction
Particle shape	Fractal	Fractal and Spherical	Fractal	Fractal	extinction, number density
Charge to radius ratio ($e^-/\mu\text{m}$)	15	5, 10, 15, 20	15	7.5	size, number density, phase function
Rain out lifetime (years)	inf. (no rain)	50, inf.	NA	inf.	extinction, number density in troposphere
Fractal dimension	2	1.5 - 3	2	2	wavelength dependence of optical depth
Monomer size (nm)	50	50	50	50	extinction

by the number density times the coagulation kernel of the effective aerosol size). A contour plot of the coagulation kernel is shown in Appendix A.1. This time does not represent a removal of the particles from the atmosphere. However, if it were much different from the removal time, the particles would be expected to reduce their number and increase their size. Figure 2.3 shows a vertical profile of the lifetime of the aerosols in the atmosphere from rainout, coagulation, and sedimentation in our best guess simulation. Sedimentation dominates the lifetime of the aerosols above 500 km, suggesting these aerosols are slow to fall. In the bulk of the atmosphere coagulation is most important in removing particles. Near the surface, the rainout is also an important removal mechanism of aerosols in this model. Of course, atmospheric motions can also impact the aerosols as discussed further below. The dynamical lifetime of the aerosols is difficult to know since we do not have an

eddy transport term. However, we do know the total lifetime of the aerosols and it is not that different from the sedimentation lifetime.

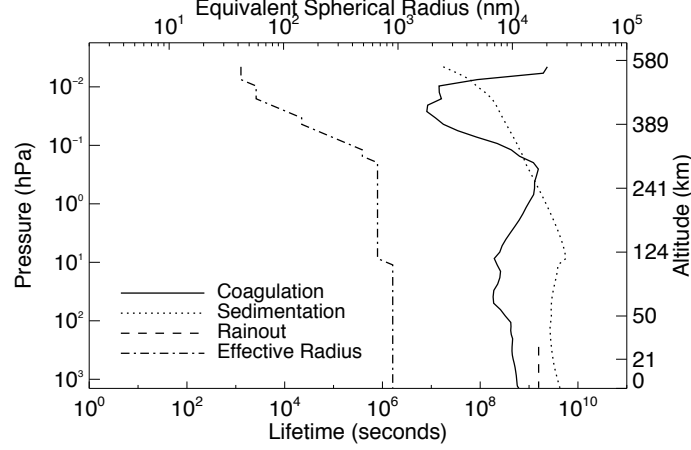


Figure 2.3: The lifetimes of the aerosols in Titan’s atmosphere due to sedimentation (dotted line), coagulation (solid line), and rainout (dashed line) for the base case at the Huygens landing site. The dashed dot line is the effective aerosol radius at each altitude at the Huygens landing site.

We calculate the total aerosol lifetime in the atmosphere by dividing the total mass of aerosols in the atmosphere by the production rate. Assuming the mass of aerosols in the atmosphere has reached steady state, this lifetime represents the average time it takes a unit mass to deposit onto the ground after being input at the top of the atmosphere. This assumption is probably false for some of our simulations, however the best guess simulation has a lifetime of 1/2 the simulation time, so it is a good assumption for our best fit parameters. The lifetime for our base case simulation is 395 years, and our other simulations range from 120-470 years (Table 2.3). The aerosol lifetimes depend most strongly on the size of the particle. Larger particles fall faster and have a shorter atmospheric lifetime. Increasing the charge to radius ratio increases the lifetime by limiting particle growth. Starting with larger particles at the top of the model decreases the atmospheric lifetime. The input mass flux

had an inverse effect on aerosol lifetime, with a larger mass input having a shorter lifetime because larger particles fall faster in the lower atmosphere. The rainout reduced the overall lifetime by 70 years. Although rainout only removes aerosols in the lowest 30 km of the atmosphere, its effects extend higher due to mixing.

Table 2.3: Aerosol lifetimes

Parameter change from base case	base case	5 $e^-/\mu\text{m}$	10 $e^-/\mu\text{m}$	20 $e^-/\mu\text{m}$	0.5x10 ⁻¹⁴ g cm ⁻² s ⁻¹	1x10 ⁻¹⁴ g cm ⁻² s ⁻¹
Lifetime (yrs)	395	120	270	470	478	469
Parameter change from base case	5x10 ⁻¹⁴ g cm ⁻² s ⁻¹	50 yr rain-out	2 nm input	660 nm input	no zonal forcing	best fit case
Lifetime (yrs)	335	327	410	344	373	257

2.3.2 The impact of particle shape on optical depth

Figure 2.4 compares the simulated wavelength dependence of the aerosol optical depth in our best guess simulation and a simulation with spherical particles. The simulation data is from the time and location of the Huygens landing site (-10° latitude and 165° longitude) for comparison with DISR observations. We also compare with ground based measurements at different seasons. We normalize the simulation results and the DISR optical depths above 80 km from Tomasko et al. [2008b] at a wavelength of 1 μm to remove any confusion from the magnitude of the optical depth and to get a more direct comparison of slopes. The wavelength dependence of optical depth depends quite heavily on the size and shape of the aerosols. Fractal particles do a much better job of fitting the observed steep slope than do spherical particles, which produce almost no slope for the sizes found in our model. That fractal particles are needed to obtain a slope comparable to observations is a very robust result and depends very little on the model parameters that we investigated. The strong

wavelength dependence at visible and near infrared wavelengths for fractals with sizes near $1\text{ }\mu\text{m}$ as compared with similar sized spheres is consistent with simulations by Wolf and Toon [2010] and Rannou et al. [1995].

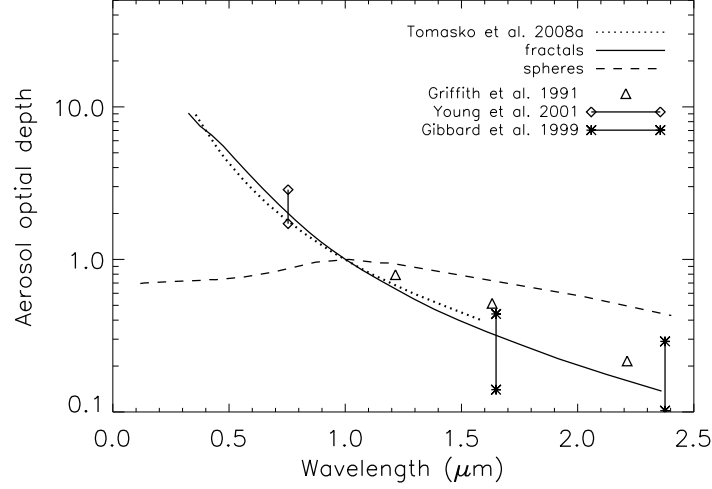


Figure 2.4: Comparing the wavelength dependence of optical depth between spherical and fractal simulations with particle sizes near $1\text{ }\mu\text{m}$. The aerosol optical depth is normalized to one at $1.0\text{ }\mu\text{m}$. Fractals with a dimension of 2 are necessary to match the wavelength dependence of the optical depth.

Figure 2.5 illustrates the magnitude and wavelength dependence of the aerosol optical depth at different altitudes from the base case model defined above. The magnitudes of our optical depth in our best guess simulation is too low at 80 km, and consistent with DISR data below that. The slope of the wavelength dependence of the optical depth is consistent between our model and DISR data at 80 km. Below this altitude, the slope of our wavelength dependence becomes too steep, our optical depths are too high in the visible and too low in the infrared. The slope of the optical depth depends on the size of the particles, their fractal dimension, and their optical properties. The flatter slopes are consistent with a higher fractal dimension or larger monomer size than used in the model. These results

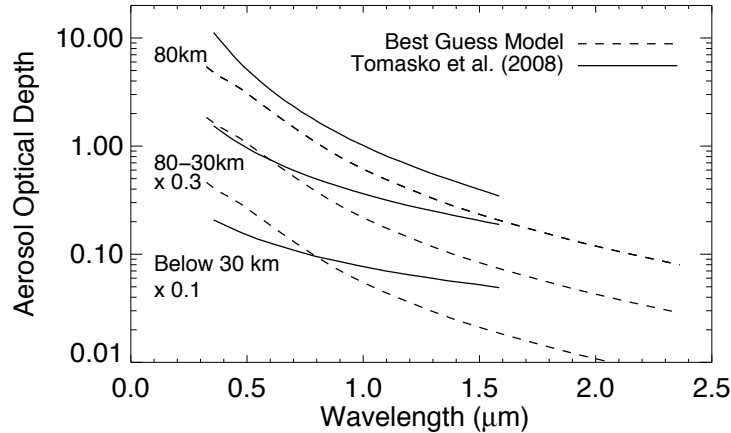


Figure 2.5: Aerosol optical depth at three different levels in the atmosphere; 80 km, 80-30 km ($\times 0.3$), and 30 km ($\times 0.1$). The simulated and observed lower altitude optical depths have been multiplied by factors of 0.3 and 0.1 to avoid overlaying points.

indicate that the nature of the particles below 80 km is changing relative to those at higher altitudes, probably due to condensation of organic gases on the aerosols near the tropopause [Tomasko et al., 2008b]. Tomasko et al. [2008b] shows that the single scattering albedo also increases below 80 km. This change in single scattering albedo indicates that the indices of refraction are also changing which is consistent with condensation of a non-absorbing material onto the aerosols. Our present model does not include condensation onto aerosols, however, we believe it is the most likely source of the discrepancies between our model and the observations below 80 km. In appendix A.4 we explore how the fractal dimension can affect the wavelength dependence of the optical depth and suggest that including a vertical dependence of the fractal dimension can improve aerosol models.

2.3.3 Effect of aerosol parameters on extinction and optical depth

In this section we compare vertical profiles of extinction and optical depth with observations from Voyager, Cassini, and the Huygens probe in order to determine the best

fit microphysical parameters as well as understand the sensitivity of each parameter. We rely heavily on the Huygens probe data for understanding the aerosol properties in Titan's lower atmosphere. However, the data from the Huygens probe is from one location in the atmosphere at one particular time. Using this data, or any of the satellite datasets, to constrain the aerosols microphysics is complicated due to horizontal gradients in the atmosphere. Nonetheless, it can be informative.

In Fig. 2.6 we plot vertical profiles of aerosol extinction (top) and aerosol optical depth (OD) (bottom) from our base case simulation and those with a mass production rate from $0.5 - 5 \times 10^{-14} \text{ g/cm}^2/\text{s}$. I have also included our best fit aerosol case (red lines). The aerosol extinction and OD have a strong dependence on the production rate with higher production rates having higher extinctions as expected. A production rate of $0.5 \times 10^{-14} \text{ g/cm}^2/\text{s}$ appears to best fit both the upper atmospheric observations from Voyager and Cassini as well as the lower atmospheric observations from Huygens. However, as we will see in the following figures, production rate is not the only property affecting the extinction and optical depth. The base case production rate of $3 \times 10^{-14} \text{ g/cm}^2/\text{s}$ is clearly too high, however. It should be noted that the Cassini observations from West et al. [2011] are at 342 nm compared with the extinction we are reporting at 525 nm. To correct for this we have multiplied the West et al. [2011] extinction by the ratio of our aerosols extinctions at 525 to 342 nm as seen in Fig. 2.4 above. This factor is 0.5. This scaling reveals the consistency between the Voyager and Cassini observations of extinction near 500 km and near 300 km.

Other aerosol parameters affect the extinction and OD in addition to the mass production rate. One is the charge to radius ratio, whose effects are seen in Figure 2.7. As will be better demonstrated in the next section, the charge to radius ratio controls the growth of the particles. Larger charges inhibit coagulation keeping the particles small and abundant, while a small charge allows for efficient coagulation causing fewer large particles. A smaller charge on the particles allows for efficient coagulation which increases extinction aloft, but decreases extinction near the surface where the larger particles are removed faster due to

sedimentation. Our best guess simulation (red lines) uses a charge to radius ratio on the particles of 7.5 electrons per micron. Our best guess simulation has a lower optical depth than the Huygen’s probe at 140 km. The Huygens data suggests an optical depth of 2, while our simulation finds an optical depth of 1 around 525 nm. Yet, our optical depth in this simulation is within the range of the stratospheric observations, and greater than two of them above 250 km. Assuming the retrievals by the spacecraft instruments were correct, this suggests that either Titan’s stratosphere had an exceptionally thick stratospheric optical depth during the Huygens descent or that our model is missing aerosols between about 140 and 250 km.

Another property affecting the extinction and OD is the removal of aerosols in the lower atmosphere. Several clouds have been observed on Titan, some apparently producing a darkening of the surface that has been interpreted as methane rain reaching the surface [Turtle et al., 2011, Griffith et al., 2012]. We include a removal term in some of our simulations in the lowest 30 km. We call this rainout but note that any removal mechanism in the lower atmosphere, such as changes to the aerosol fall velocity due to compaction or condensation could also have the same effect. Comparing the base case (yellow) and a case with rainout (red) in figure 2.8 (top), we see that the effect of the rainout rate on the vertical extinction is to greatly reduce tropospheric extinction near the surface. We conclude that the shape of the extinction profile at low altitude is not consistent with removal by precipitation in which aerosols are removed from a high altitude and quickly deposited on the surface. However, the shape of the profile is consistent with sedimentation towards the ground where the particles are ultimately removed onto the surface.

In Figure 2.8 we also explored the effect on our simulations of inputting different sized particles at the top of our model. There is very little difference between inputting 2 nm and 50 nm particles at the top of the model. However, increasing the input particle size to 660 nm enhances the extinction above 350 km. This enhancement is inconsistent with the observations and suggests that the aerosols do not coagulate into large aggregates above 500

km. Below 350 km the particles in all simulations have coagulated to the size allowed by their charge to radius ratio and their residence time. The altitude of methane destruction and peak aerosol production is somewhat debated. However, the consistency with which our extinction matches the observations to the top of our model suggests the bulk of haze production occurs at high altitude, above 500 km. Figures 2.8 and 2.7 have an extended haze layer in those simulations in which the particles have reached a size of $0.5 \mu\text{m}$ or larger by 350 km. A more thorough discussion of the extended haze is in preparation.

As discussed below the extinction profile at the Huygens landing site is also affected by the meridional distribution of the aerosol. The ad hoc torque used to accelerate the zonal winds reduces the meridional aerosol gradient, which in turn increases the optical depth in the tropics relative to a case without the torque, especially between 50 km and 200 km (which curiously is a region in which the torque is not applied). As we will see in section 3.5, the aerosols tend to accumulate at the poles. If we reduce the meridional gradients however, more aerosols will be in the tropics where the Huygens probe descended.

2.3.4 Particle size and number density

The charge to radius ratio in our model has the largest effect on the particle size and number density. The phase functions are strongly dependent on the size of the aerosols, with large aerosols being more forward scattering. They are also dependent on the fractal dimension with more spherical aerosols being less forward scattering. However, the phase function does not depend on the amount of aerosol in the model. Our phase functions at 100 km for simulations with varying charge to radius ratio are compared to the DISR data from Tomasko et al. [2008b] in Figure 2.9. The data at successive wavelengths are multiplied by consecutive orders of 2 (2,4,8,...) in order to separate the data and make the plot more readable. We normalized all the phase functions so that the area under the curve is unity before multiplying them by successive orders of two. To match the phase functions reasonably well, we needed to charge the particles in Titan's atmosphere. At 100 km our

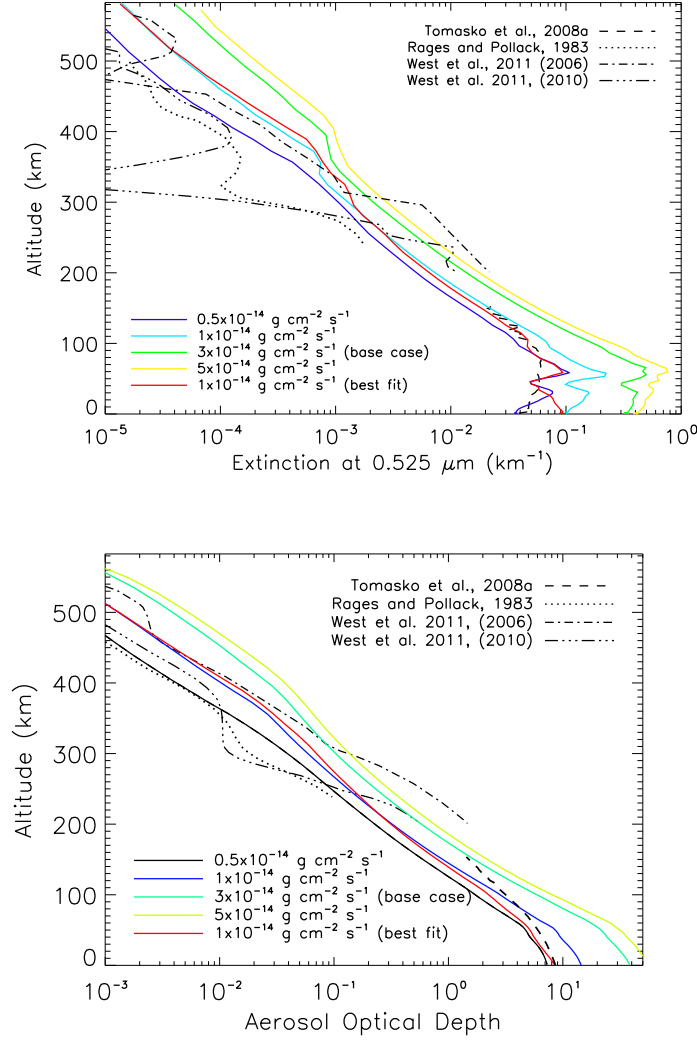


Figure 2.6: Top) Extinction as a function of altitude at 525 nm for several different mass inputs compared with observations from Voyager (500 nm) , Cassini (342 nm) $\times 0.5$, and Huygens (530 nm). Each colored line corresponds to a model simulation and is labeled with the mass input at the top of the model. Bottom) Aerosol optical depth as a function of altitude with the same parameters as the extinction plot.

best fit had a charge to radius ratio of $7.5\text{-}10 \text{ e}^-/\mu\text{m}$. Charge to radius ratios of 5 and $15 \text{ e}^-/\mu\text{m}$ were well below and above the observations at most wavelengths. We chose $7.5 \text{ e}^-/\mu\text{m}$ as our best guess simulation because it is more consistent with the number density data as

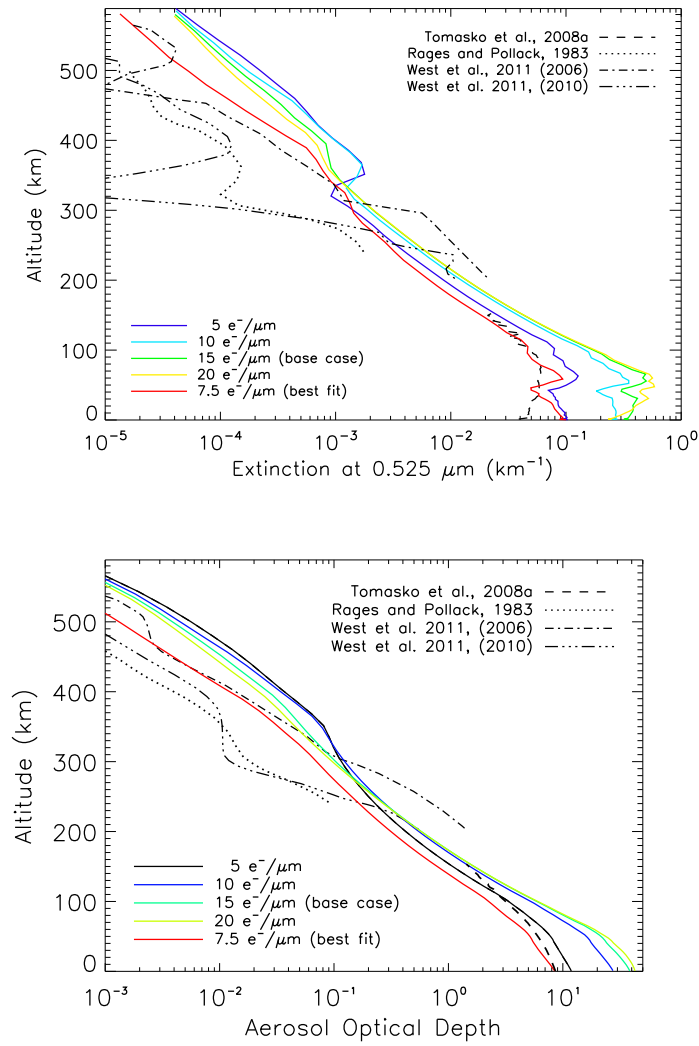


Figure 2.7: Same as Fig. 2.6 except we are changing the charge to radius ratio on the particles.

well.

Figures 2.10 and 2.11 display the size distribution of the aerosols. Figure 2.10 shows that the effective radius of the aerosols grows quickly because of coagulation between 400 and 350 km and is fairly constant below that. The runs have systematically smaller particles with increased charge to radius ratio, because increased charging inhibits particle growth.

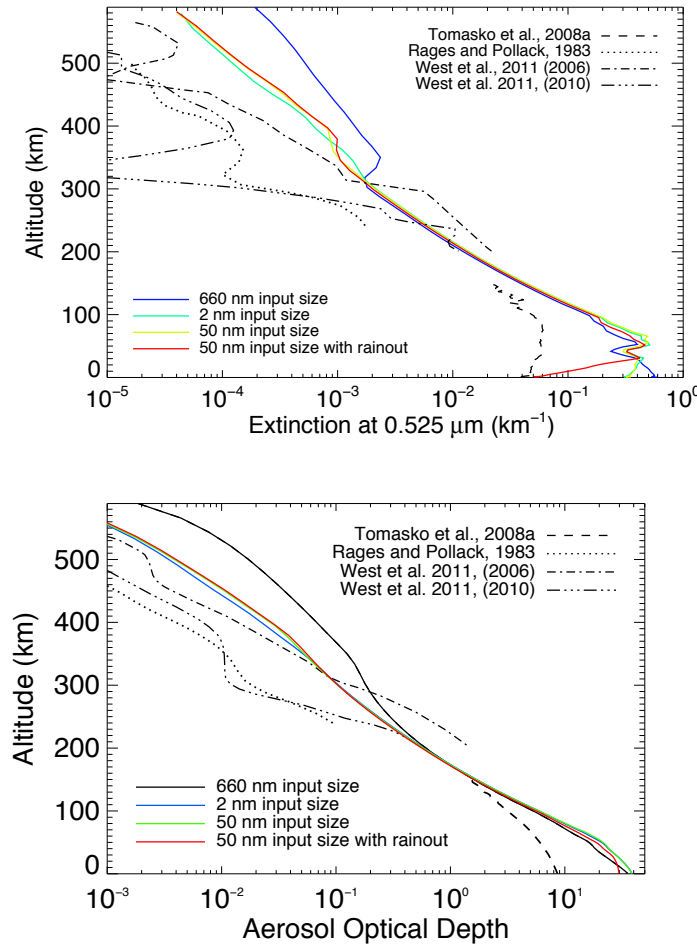


Figure 2.8: Same as Fig. 2.6 except that we are comparing rainout with a 50 year lifetime and particle input sizes in our base case model.

Our best guess particle sizes are slight smaller than those of Lavvas et al. [2010] between 100 and 300 km and consistent with their sizes below 100 km. Figure 2.11 gives the size distributions of the particles in the atmosphere at several different altitudes from our base case simulation. The peak of the size distribution of the aerosols gets larger lower in the atmosphere in the base case, which is expected from coagulation.

Comparing figures 2.10 and 2.12 we see the aerosol particle size and number density are

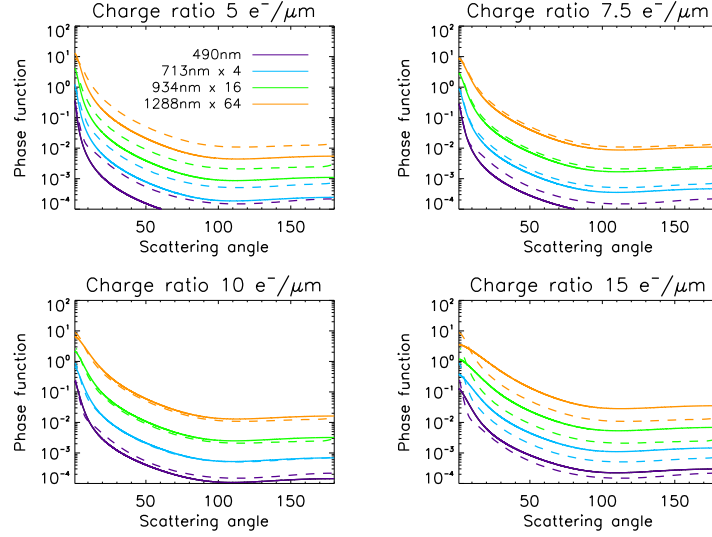


Figure 2.9: Phase functions at 100 km from simulations with a varying charge to radius ratio of 5-15 $e^-/\mu\text{m}$. Solid lines are the model results; dashed lines are from Tomasko et al. [2008b] DISR data. The base case simulation (lower right) has a charge to radius ratio of 15 $e^-/\mu\text{m}$. Once normalized, the curves are multiplied by consecutive factors of two to separate out the data to make each line visible.

inversely proportional. Figure 2.12 compares our simulations of the aerosol number density for cases with a varying charge to radius ratio to those of Lavvas et al. [2010] and observed aerosol number densities from Tomasko et al. [2008b] and Vinatier et al. [2010]. The decrease of number density between the top of the model and 300 km is expected from coagulation and is seen in our simulations and those of Lavvas et al. [2010], however observations do not extend to high enough altitudes to show this trend. The modeling results suggest a high altitude production and coagulation region for the aerosols, and that the particles have reached aggregate sizes by 350 km. Below 350 km, our best guess simulation aerosol number density continues to increase towards the surface. Our results are very consistent with both the Lavvas et al. [2010] model results and the observations from Cassini and Huygens. Although, optical data generally have trouble retrieving number densities because many aerosol properties including size need to be assumed, the remarkable consistency between

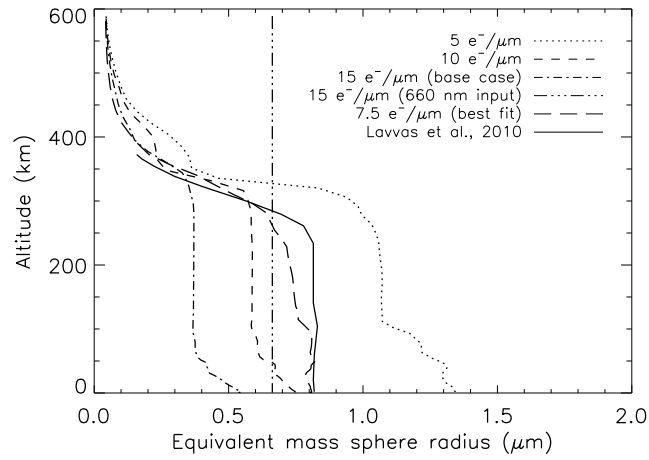


Figure 2.10: Vertical profile of the effective radius of the aerosols from our simulations compared with the simulations of Lavvas et al. [2010].

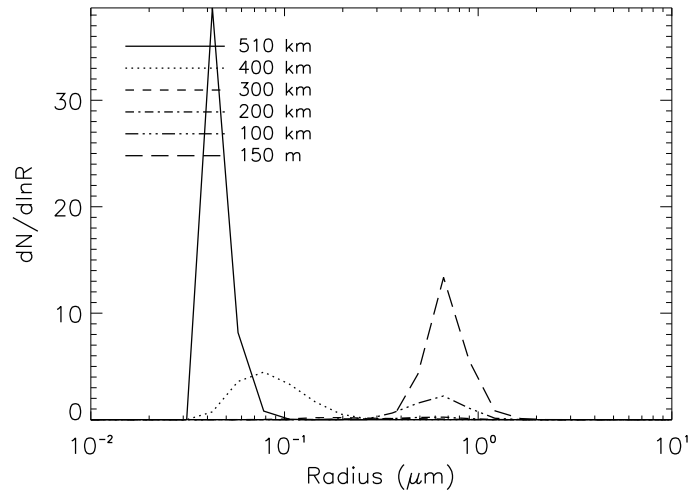


Figure 2.11: The size distribution of our best guess case aerosols at the Huygens landing site at six different levels in the atmosphere. The radius is the equivalent mass sphere radius.

the Vinatier et al. [2010] data and the Tomasko et al. [2008b] data at 150 km lends credence to these derived properties. The model of Lavvas et al. [2010] produces aerosols much higher than our model, however their particles coagulate into aggregates at the same altitude, about 300 km.

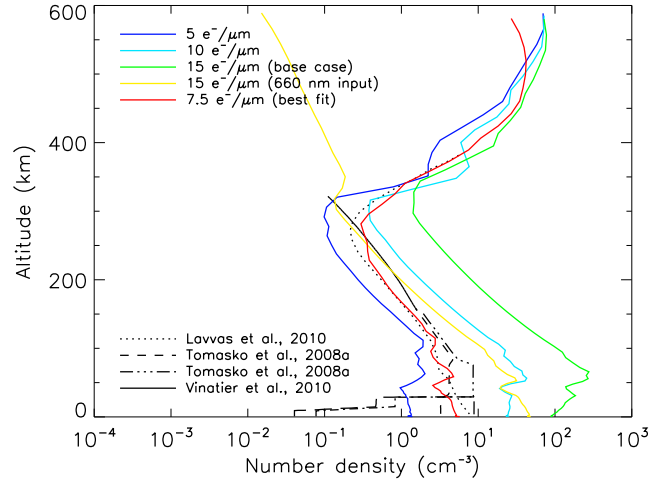


Figure 2.12: Simulations varying charge to radius ratios and particle size compared with retrieved number densities.

From these sensitivity studies we can constrain our best guess aerosol parameters. The comparisons with data suggest a mass production rate of about $1 \times 10^{-14} \text{ g cm}^{-2} \text{ s}^{-1}$ and a particle charge to radius ratio of $7.5 \text{ e}^-/\mu\text{m}$ will produce the best fit between simulations and observations. Inputting small particles of 50 nm or less at 580 km altitude produces simulations that are consistent with the observations of extinction in the upper atmosphere. It does not appear that rainout is necessary to simulate the vertical extinction profile in Titan's troposphere, and may be inconsistent with observations.

2.3.5 Aerosol latitudinal distribution and seasonal cycle

The extinction in our simulations exhibits weak latitudinal gradients (Figure 2.13), however there is an enhancement of aerosols at the poles compared to the tropics below 100 km. We also see an enhancement of aerosol extinction above the winter pole between about 100 and 300 km. This can be interpreted as a polar hood. Lorenz et al. [2006a] observed a seasonally changing aerosol polar hood, which peaks during spring. In our model the polar aerosol enhancement is greatest in the fall-winter hemisphere. Our model, as further discussed below, has many features that are out of phase with the observations, but otherwise similar to the observations.

All of our simulations exhibit this same basic structure of enhanced aerosols at the poles, especially over the winter pole. These features are robust and the microphysical parameters have small effects on the haze distribution. However, the ad hoc equatorial torque, mentioned in section 2.1, acts to smooth out the meridional gradients, especially below 200 km. The runs without this ad hoc torque have gradients at 50 km up to a factor of 8 in aerosol extinction between equator and pole, compared with 2-4 in our forced models. The latitudinal gradients in our unforced model are more consistent with those of Lebonnois et al. [2009]. Their model has latitudinal gradients of a factor of 10 or more. Our model produces an extended haze layer at 350-400 km coming off of the top of the polar hood at a similar altitude as seen in Voyager data [Rages and Pollack, 1983] and by West et al. [2011]. This phenomenon is affected by the aerosol parameters chosen, yet robust against the addition of the ad hoc torque. We discuss the extended haze layer further below.

Figure 2.14 shows the latitudinal profile of optical depth at 100 km measured by the VIMS instrument on Cassini [Rannou et al., 2010] compared with our best guess simulations at different seasons. The observations are near LS 299. Both data and simulations are consistent in magnitude and indicate the northern latitudes have larger optical depths than the southern latitudes at this time of year. However, our simulated optical depths display an

enhancement at the poles relative to the equator, which is opposite of the observations. The polar region with the highest aerosol concentration switches on a seasonal timescale in our model, however the tropics always have lower concentrations of aerosols. This can be seen clearly in the colored lines in Fig. 2.14 which represent different times of year. Changing the microphysical parameters or horizontal resolution has very little effect on these results. Simulations without the ad hoc torque have larger polar enhancements than the best guess simulation shown.

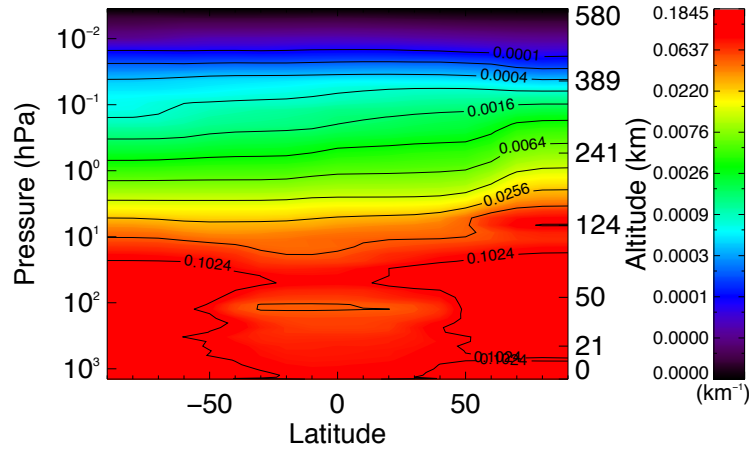


Figure 2.13: The meridional distribution of the aerosol extinction (km^{-1}) in our best fit model at 525 nm during the northern winter.

The haze mass mixing ratio in our best guess simulation is similar in magnitude with that derived by Vinatier et al. [2010] from Cassini/CIRS infrared spectra of Titan's limb during the northern winter in the altitude region between 140 and 480 km (Fig. 2.15). However, our mixing ratio peaks at the winter pole while the CIRS data peaks in the tropics, which could be more evidence of our model phase lag.

The latitudinal distribution of the aerosols in Titan's atmosphere varies seasonally, with the polar hood oscillating back and forth from pole to pole. Figure 2.16 compares

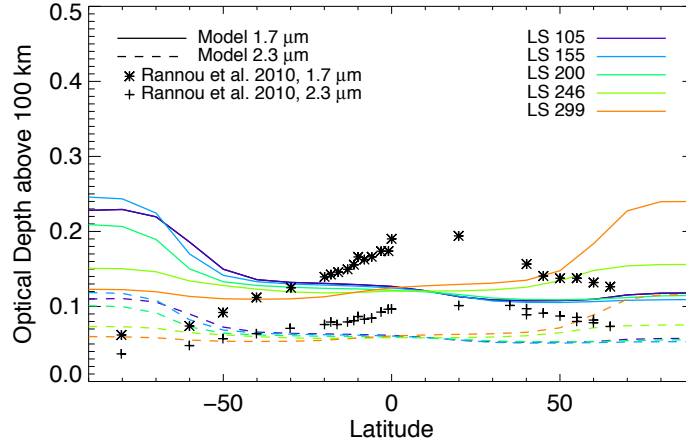


Figure 2.14: The latitudinal profile of optical depth at $1.7 \mu\text{m}$ and $2.3 \mu\text{m}$ above 100 km from our best fit simulation compared with data from Rannou et al. [2010]. LSS 299 is during the Huygens probe descent.

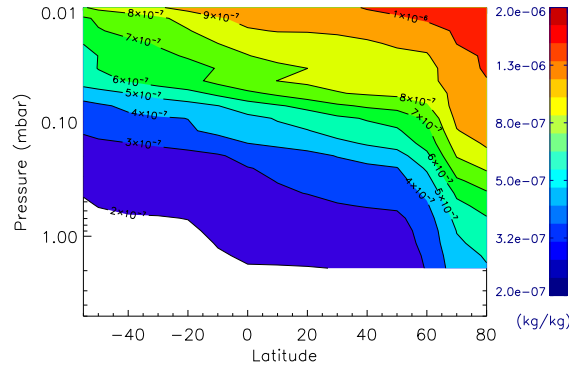


Figure 2.15: The haze mass mixing ratio for a portion of the stratosphere from our best guess simulation during the northern winter. The color scheme and region of the atmosphere were chosen to better compare with Vinatier et al. (2010, Fig. 13).

Titan's aerosol area density ($\mu\text{m}^2/\text{cm}^3$) at LS 155, 200, 246, and 299 (upper left moving counter clockwise) in our best guess simulation. As our simulation progresses throughout

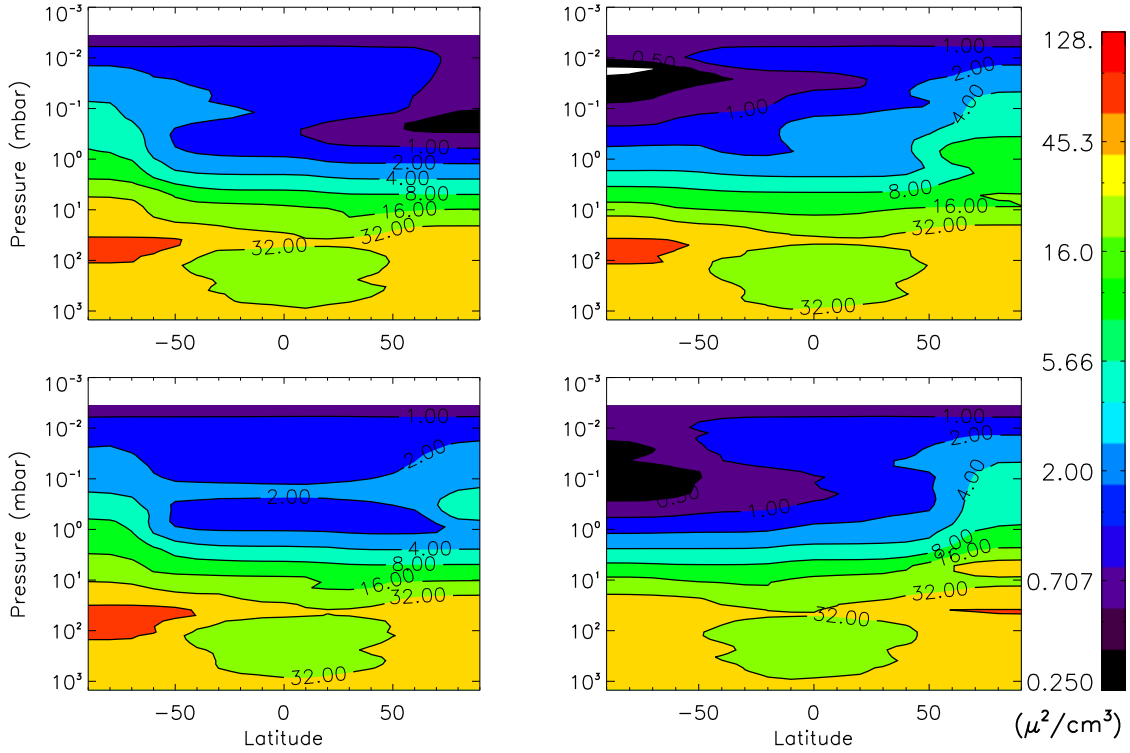


Figure 2.16: The area density ($\mu\text{m}^2/\text{cm}^3$) of Titan's aerosols at LS 155, 200, 246, and 299 (upper left moving counter clockwise) in our best guess simulation.

the year, you can see the polar hood in the South shift to the North. The extended haze layer projects off of the polar hood. Around equinox (lower left) the haze layer is symmetrical and at a lower altitude than at the solstice. The summer hemisphere on Titan has ascending air while the winter hemisphere has descending air in the stratosphere. The upper stratosphere has meridional winds from the summer pole to the winter pole. This overturning Hadley cell has been well documented by monitoring the abundance of trace gas species [Teanby et al., 2012, Bampasidis et al., 2012]. From this understanding, we can explain the extended haze layer in the model. The term extended haze layer is a bit misleading, since there is not really a high concentration of haze in this layer, but instead a gap in the haze below

this layer. This is clear in West et al. [2011] where they plot the haze extinction profile next to an exponential fit. In our model the aerosols are suspended and removed in the summer hemisphere stratosphere and move towards the winter pole. This gap in the aerosols is most prevalent where they are being removed. The winter hood forms as aerosols are transported to that region from other latitudes and the production region above. The altitude of the extended haze layer then is determined from a balance between the dynamics and the fall velocities of the particles. A more thorough description of the movement of the extended haze layer will be explored in a later paper, however it appears that the overturning Hadley cell determines the altitude and affects the strength of the haze layer.

2.3.6 Effect of haze on Titan’s albedo

We calculate the geometric albedo using the GCM radiative transfer scheme which computes the hemispheric albedo. Therefore, wavelength resolution is low, and we used a factor of $2/3$ to crudely convert the hemispheric albedo to the geometric albedo. The $2/3$ comes from the assumption of isotropic scattering by Titan. These assumptions likely affect the comparison in Figure 2.17. Our best guess model has a reasonable match to the geometric albedo considering our assumptions.

Lockwood et al. [1986] and Lorenz et al. [1999] use Voyager, Hubble, and ground-based data to illustrate the seasonal cycle of Titan’s geometric albedo and the seasonal cycle of the north/south asymmetry of the albedo. Toon et al. [1992] proposed that the changing balance between aerosol settling and winds was responsible for the redistribution of aerosols that cause the seasonal and wavelength dependent albedo variations. Hutzell et al. [1996] elaborated on that idea with a 2-D model that was not radiatively interactive. Hutzell et al. [1996] showed that at a given altitude, the particles suspended there should be smaller in the winter hemisphere and larger in the summer hemisphere due to atmospheric motions sometimes lofting the particles (making them larger), and sometimes carrying the particles downward (making them smaller). Rannou et al. [1995] further showed that these mecha-

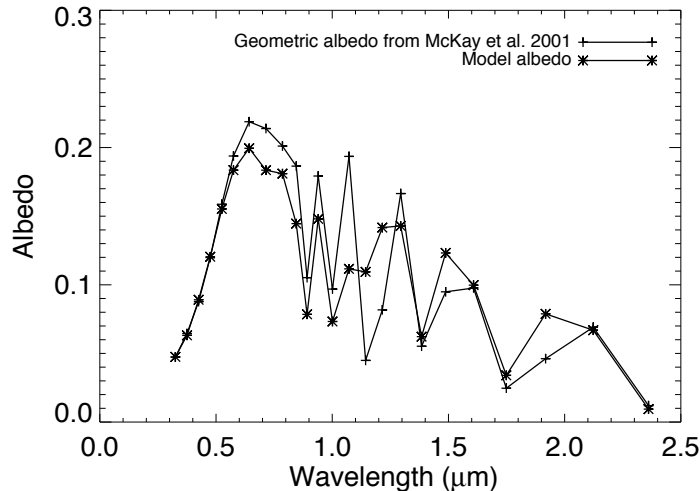


Figure 2.17: Titan’s geometric albedo from the best fit model compared with data from McKay et al. [2001].

nisms were amplified in models in which the microphysics and dynamics interact through the radiation field. Our model reproduces this effect, as seen by the seasonal shift in the aerosol area density (Fig. 2.16), number density, and effective radius. This is a very robust outcome of the model and the gross features are consistent in all simulations. Similar to our previous results in Figure 2.10 and 2.12, we find an inverse relationship between number density and effective radius as the seasonal cycle of the winds modulates the aerosol properties. Where the particles are being suspended, the effective radius increases and the number density decreases.

Our model supports the notion that the effect of the seasonal changes in aerosol effective radius and number density on Titan’s albedo is wavelength dependent. Figure 2.18 (top) shows the seasonality of the wavelength dependence compared with observations compiled by Lorenz et al. (1997). The wavelength dependence is expected to shift between greater than one and less than one as the seasons change. This means that the bright hemisphere

is expected to shift seasonally. Below 700 nm, our model makes this shift with the southern hemisphere being brighter before the autumn equinox. After the autumn equinox, the northern hemisphere is brighter. The observations are nearly contemporaneous with our simulations at LS 180, the asterisk symbols in the Fig. 2.18 (top). At this time, our model should be brighter in the northern hemisphere at wavelengths below 700 nm, however we do not achieve the observed ratios until about LS 210, indicating our model may be phase shifted. However, albedos at the longer wavelengths in our model are not phase shifted with respect to the observations.

Our model produces a seasonal N/S albedo at 475 nm that is similar in magnitude to the observations, however it is phase shifted by about a season from the observations (Figure 2.18 bottom). We previously noted that our polar hood is forming about one season too early. Due to the overturning Hadley cell, the winter hemisphere has a higher abundance of aerosols. At blue wavelengths, these aerosols should be more absorbing, thus the winter hemisphere should have a lower albedo. The observations show that the maximum northern hemisphere albedo is during the northern autumn, about 80 degrees out of phase from the solar cycle [Lorenz et al., 1999]. This phase lag in Titan's atmosphere is due to the dynamical time constants at the altitude of the main haze layer being about one season [Flasar et al., 1981]. Our model shows the highest northern hemisphere albedo at LS 260 degrees, near the southern summer solstice, so our model is phase shifted by 170 degrees from the solar cycle and about 90 degrees from the observations. We suggest caution from pulling strong conclusions from this discrepancy, however. The geometry from the plane parallel radiative transfer in this model does not directly compare to the geometry behind the observations of the geometric albedo.

2.3.7 Aerosol radiative effects

The heating and cooling rates averaged over an Earth year are compared between our best guess model and Huygens observations by Tomasko et al. [2008a] in Figure 2.19. The

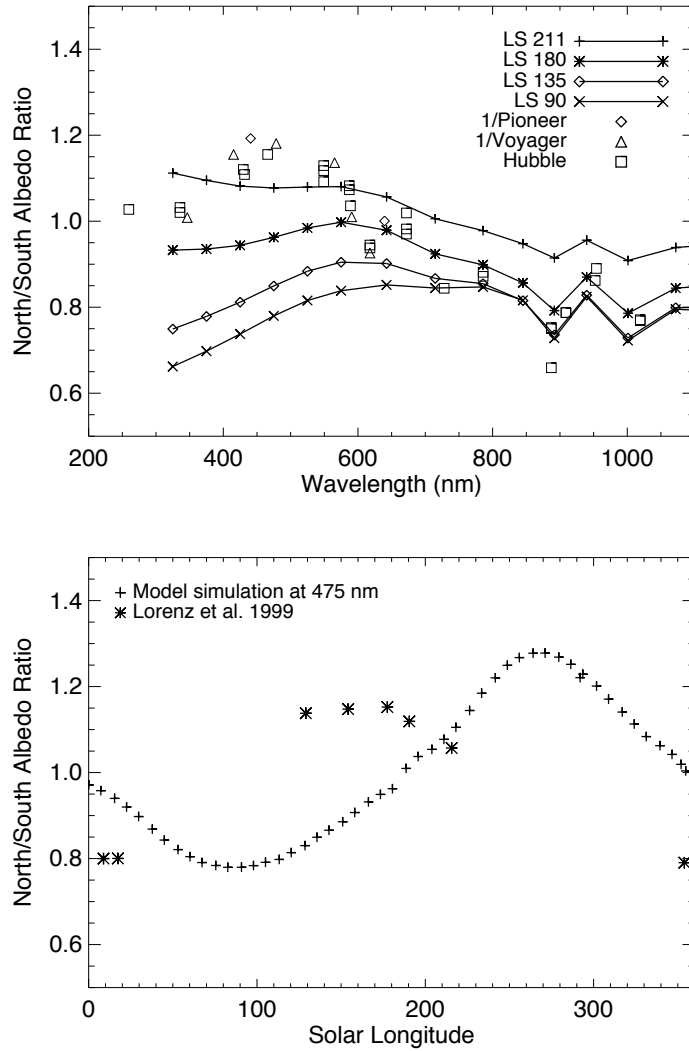


Figure 2.18: The top panel shows the wavelength dependence of Titan's north/south albedo asymmetry compared with observations reported in Lorenz et al. [1997]. Bottom panel compares the Lorenz et al. [1999] albedo ratio at 439 nm with our best fit model results at 475 nm.

heating and cooling rates in our model are strongly dependent on the aerosols and their properties. Specifically, changing the imaginary indices of refraction can affect the heating and cooling. Likewise, changing particle size or optical depth by altering the charge to radius ratio or production rate will impact heating rates. Too large of a production rate

leads to an excess in heating and cooling above the tropopause, because the optical depth is too high, and a dearth below the tropopause compared with observations. The higher heating than cooling at all altitudes in our model suggests that heat is being transported away from the tropics at all altitudes. Our model does not include any non-LTE effects. The LTE approximation is sufficient up to 0.01 hPa, or about 500 km in our model [Yelle, 1991]. Above that altitude, our model which assumes LTE, over estimates the cooling rate up to several tenths of a degree per day. This would widen the gap between our heating and cooling rates in Titan’s upper atmosphere at this location.

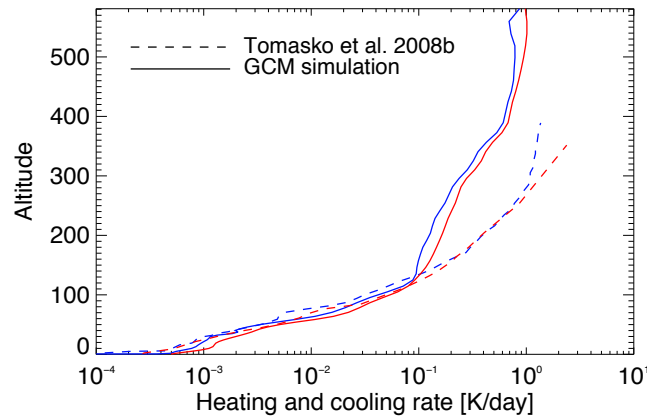


Figure 2.19: The shortwave heating (red) and longwave cooling (blue) rates at the Huygen’s landing site during the northern winter averaged over one Earth year from the best fit simulation.

Generally, our model produces temperatures that are about a few Kelvin lower than observations at the surface, about 10 K warmer just above the tropopause, and about 20 K cooler than observed above 300 km (Fig. 2.20, top). The top panel is our best guess aerosol parameters and our best match to Titan’s heating profile. The middle and bottom panels are our base case aerosol parameters with and without any zonal forcing. It is interesting to note that accelerating the winds drives a temperature gradient in the model with a latitudinal

relationship that is similar to observations. The latitudinal temperature gradients in both the forced simulations and the observations switch signs around 1 hPa. In our best guess simulation (Fig. 2.20, top) the tropics are warmer by over 10 K above 1 hPa, while below that, the northern polar regions get warmer by up to 20 K during the northern winter. These gradients are similar to the observations. The temperature structure, i.e. latitudinal gradients that switch sign in the atmosphere, are related to the superrotating zonal jets. Further evidence of this comes from Liu et al. [2008], who used CAM 2.0 to reproduce 100 m/s winds in Titan’s stratosphere by forcing latitudinal temperature gradients. It seems that forcing either the winds or the temperatures to the observations causes the other to conform to the observations.

Aerosol microphysical and radiative parameters can have an impact on the temperature profile. By reducing the infrared absorption coefficients from Vinatier et al. [2007b] by a factor of 3 we get a temperature increase in the upper atmosphere up to 20 K, making our model consistent with the observations. Also changing the charge to radius ratio and mass production rate at the top of the atmosphere affects the temperatures. Low surface temperatures in this model are the result of excess aerosol extinction in the upper atmosphere blocking light from reaching the surface.

2.3.8 Dynamical implications

A robust result from our model is producing rising motion in the summer hemisphere and descending motion in the winter hemisphere aloft (Figure 2.21, top). Both of these results are consistent with a pole-to-pole Hadley circulation first produced by Hourdin et al. [1995] and produced under all parameters explored in our model. The vertical winds in the stratosphere are comparable to or stronger than the particle fall velocities in our model. Thus the vertical aerosol transport is dominated by these vertical winds as opposed to sedimentation. The meridional winds travel from the summer to the winter hemisphere aloft and are very strong, several meters per second, near the top of the model (Figure 2.21,

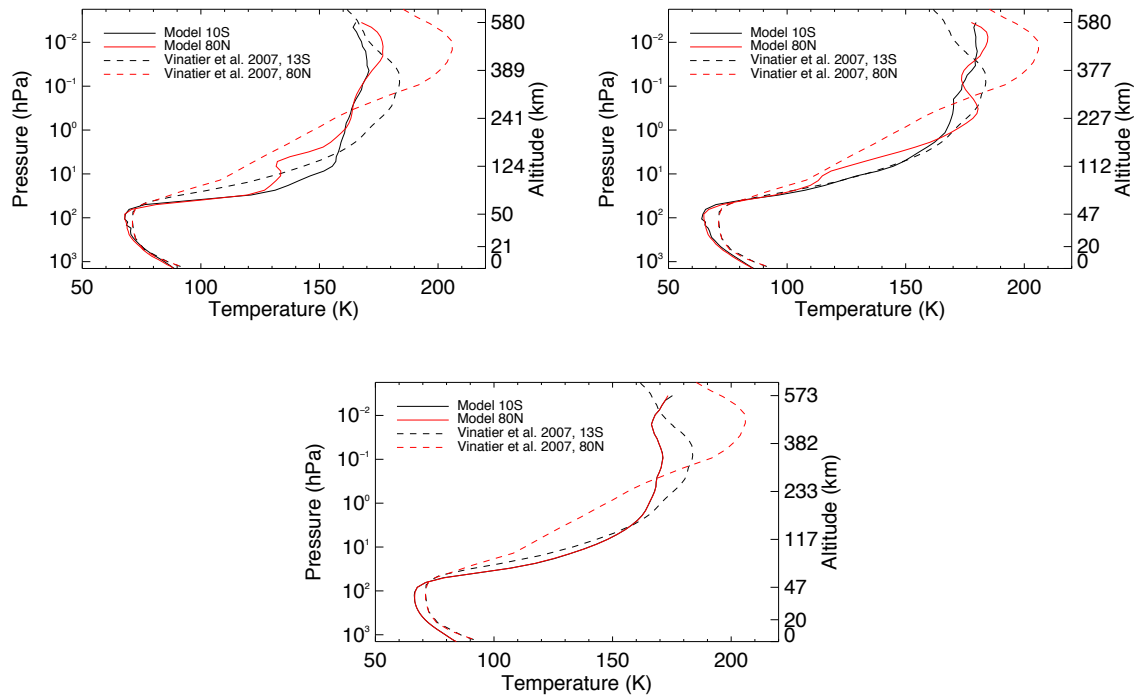


Figure 2.20: Comparing temperature profiles from the best guess model (top) and base case with and without the ad hoc forcing (middle, bottom) with CIRS data from Vinatier et al. [2007a] during the same season as the Huygens landing.

bottom). During the equinox, the meridional wind weakens and we get two pole to equator cells at the top of the model creating convergence aloft in the tropics.

The large observed temperature gradients on Titan are associated with cyclostrophic balance with superrotating winds. Reproducing these temperature gradients should drive the superrotating winds. As mentioned before, our model does not reproduce the temperature gradients, and is also unable to reproduce the 100+ m/s zonal winds observed in Titan's stratosphere. Therefore we implemented an ad hoc acceleration to the winds. In the low resolution model (10x15 degrees) without this ad hoc torque, the bulk of the atmosphere actually moves retrograde (westward) at speeds greater than the eastward zonal jets (Figure 2.22, top).

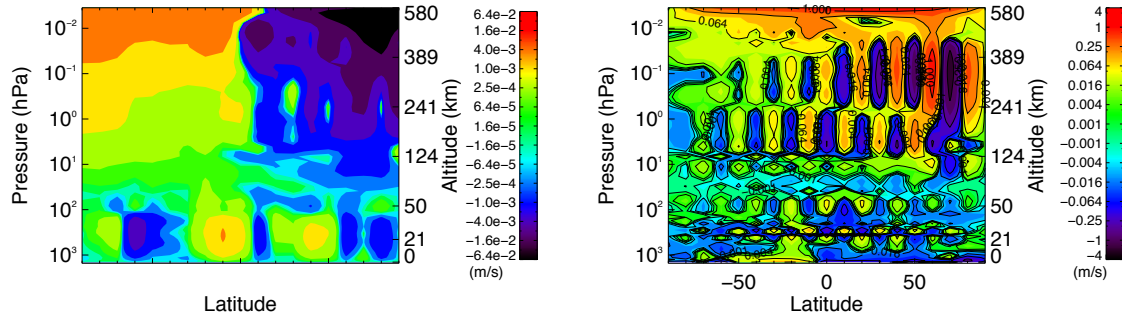


Figure 2.21: Wind fields during a northern winter in the best fit simulation. The top plot shows the vertical air motion, where warm colors are rising air. The bottom plot is the meridional winds where warm colors are winds from south to north. These winds are averaged over an Earth year at the time of the Huygens probe descent.

An inability to reproduce the superrotating winds has been a common theme among Titan GCM's. Only recently have 3D Titan GCMs been able to reproduce the superrotation. Newman et al. [2011] were able to obtain superrotation in the Planet-WRF model run at 5.5 degree in longitude and 5 degrees in latitude horizontal resolution with 54 vertical grid levels extending to 420 km altitude, though it required 75 Titan years for steady state to be reached. However, in order to do so they eliminated all the imposed horizontal diffusion in their model. If they did not remove all of the explicit horizontal diffusion they had very small latitudinal temperature gradients similar to the ones shown in our unforced model (Fig. 2.20, bottom). Newman et al. [2011] also ran a lower resolution model (11.25x10 degrees) and were able to reproduce the superrotation.

We made progress at increasing our zonal winds by moving to higher resolution, reaching wind speeds of 25 m/s in the unforced model at 4x5 degrees resolution. However, even higher resolution is required to overcome the dissipation in the model, if it is possible. We know the finite volume dynamical core used in CAM is highly dissipative at low resolution. However, it is computationally unfeasible at this time to further increase the resolution of

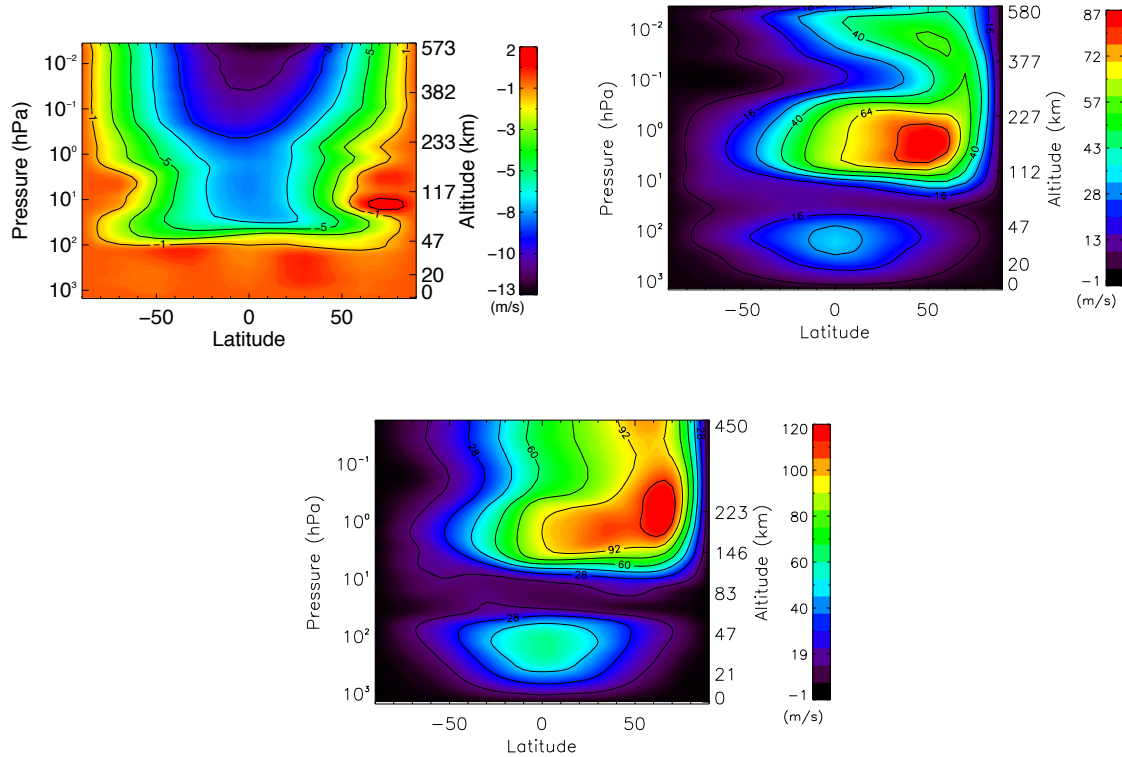


Figure 2.22: Titan’s zonal wind (m/s) from the low resolution unforced simulation (top), low resolution forced simulation (middle), and high resolution forced simulation (bottom) after 16,16, and 6 Titan years respectively. Note the difference in color bar scales.

this model. The long time constants on Titan make running this GCM computationally very expensive. It also may be the case that this dynamical core does not accurately treat angular momentum [Lebonnois et al., 2012b]. Lebonnois et al. [2012b] found that the non-physical angular momentum residuals dominated the angular momentum budget in simulations of Venus’s atmosphere using a newer version of CAM. Adding topography to the model helped increase the physical angular momentum transfer from the surface to the atmosphere in their study. We added the topographic map from Lorenz et al. [2013] to our GCM and found no changes in the zonal wind speed aloft. The topography on Titan is relatively flat, reaching

elevations less than 1 km over most the planet. Adding it to our model only changed the wind speed and pattern near the surface.

After much effort and modifications attempting to accelerate the zonal winds, we have had little success. Thus we chose to force the zonal winds with an ad hoc acceleration term. A constant forcing allowed us to reproduce not only magnitude of the zonal wind, but it's seasonal cycle in magnitude and latitudinal oscillation as well. The forced zonal wind is shown in Figure 2.22 (middle). The zonal jet oscillates from hemisphere to hemisphere peaking at about 0.1 hPa in the mid latitudes during the northern fall season with magnitudes of 140 m/s. This wind is consistent in altitude and latitude with the peak of the zonal jets reported by Achterberg et al. [2010]. The wind contours shown in Fig. 2.22 are at the time of the Huygens probe descent and near the minimum of the seasonal peak wind speed. The seasonal cycle in the model winds is consistent with the observed zonal jet, which is in the South during the southern summer and fall. The peak magnitude of the simulated zonal jets is slightly less than that of the observations (190 m/s), however that could easily be adjusted in the model by increasing the forcing or resolution. The lower than observed wind speeds are consistent with our temperature gradients being weaker than the observations as well. Running the model at a higher resolution, 4x5 degrees, increases the maximum zonal wind speed by more than 30 m/s (Figure 2.22, bottom). It should be noted that our winds in this simulation reach a maximum speed of 170 m/s. Along with our increased magnitude in wind in the high resolution simulation is a stronger latitudinal temperature gradient at the top of the model. A further increase in resolution would likely produce even faster winds.

The superrotation index (Figure 2.23) is defined as the angular momentum in the atmosphere normalized by the angular momentum of an atmosphere at rest with respect to the surface of the planet [Read, 1986]. Newman et al. [2011] found a superrotation index slightly less than 2 for the entire atmosphere and reaching 15-18 for the region between 0.009 and 2 mbar. In our low resolution unforced model the atmosphere decelerates over time with very little seasonal fluctuation. The high resolution unforced model (4 x 5 degrees)

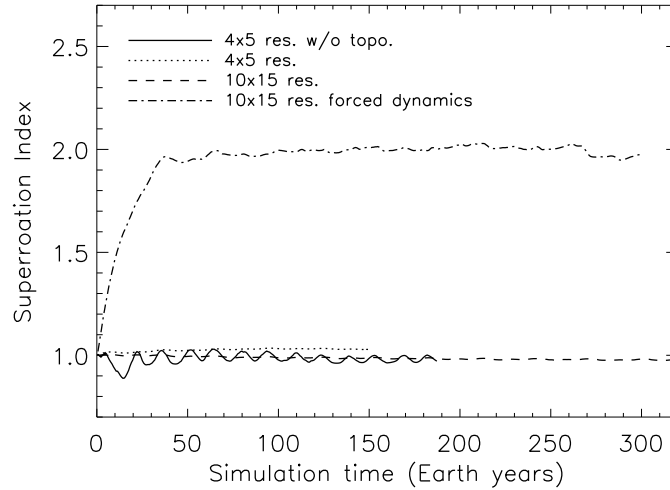


Figure 2.23: Effect of resolution and ad hoc torque on the superrotation index for Titan’s entire atmosphere.

also decelerates over time. The forced model increases its superrotation quickly over one Titan year. It reaches equilibrium near a superrotation index of 2. It is clear our non-forced model does not exhibit the superrotation in Titan’s atmosphere that other Titan GCMs have recently been able to achieve [Newman et al., 2011, Lebonnois et al., 2012b].

Another common problem for Titan GCMs is matching the tropical surface winds to the prograde wind direction inferred from the longitudinal dunes seen there. A recent hypothesis for the directions of the longitudinal dunes is proposed by Tokano [2010]. He finds prograde winds of 1.5 m/s at the equinoxes despite the weaker but much more prevalent retrograde winds (1 m/s) seen the rest of the year. His findings suggest that the threshold wind speed for lifting particles and creating dunes is between 1 and 1.5 m/s. The surface winds in our forced model are dominated by effects of topography and zonal forcing. Our winds in the forced model are prograde at some locations in tropics, however, they are not enhanced at the equinox (Figure 2.24). Our model does produce surface winds with a

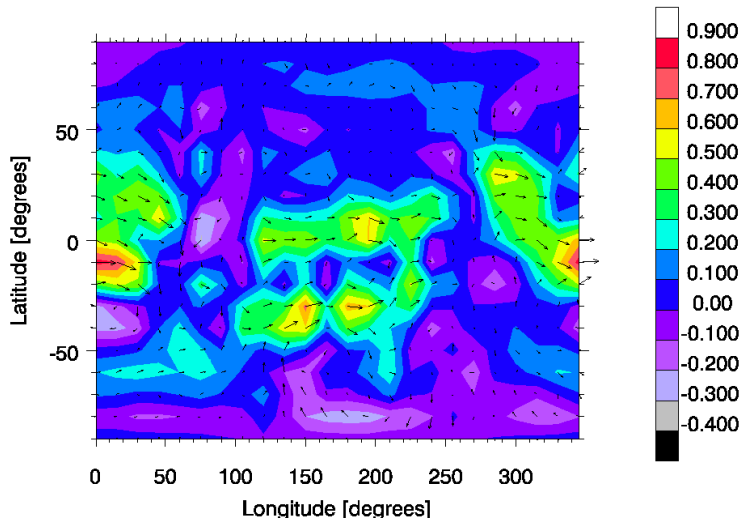


Figure 2.24: The surface winds on Titan at the autumnal equinox from our best fit simulation.

magnitude peaking near 1 m/s, which is consistent with a lack of waves on the lakes of Titan [Lorenz et al., 2010]. A more detailed study of the Titan’s surface winds with CAM is in progress.

2.4 Conclusions

We have successfully coupled the aerosol microphysical model CARMA to the Titan CAM dynamical core and radiation scheme. We modeled the aerosols as both spheres and fractals, and confirmed that the fractal particles better match observations, especially the wavelength dependence of optical depth. We identify a best fit set of aerosol parameters that are most consistent with spacecraft observations. These include a charge to radius ratio of about 7.5 electrons/micron, consistent with Lavvas et al. [2010], which is constrained by observations of aerosol number densities and phase functions. We found a mass production rate of 10^{-14} g/cm²/s best fits the extinction and optical depth in both the stratosphere and troposphere. This production rate is a factor of 3 lower than Lavvas et al. [2010], but

consistent with most pre-Cassini values reviewed in McKay et al. [2001]. The extinction in our best fit case is roughly constant in the troposphere and in agreement with Huygens data. We modeled a removal mechanism with a 50 year rainout lifetime, which led to an extinction profile that rapidly declined toward the surface contradicting the observations. We attribute the small discrepancy in the extinction profiles between our model and Huygens data to condensation in Titan’s atmosphere, which we are not modeling. This condensation softens the slope of the wavelength dependence with optical depth, which is likely due to the particles becoming more spherical or larger. Particles becoming more spherical would fall faster and be removed more quickly. We found that the softening of the slope of that wavelength dependence as seen in Titan’s lower atmosphere can be accurately modeled using fractal dimensions of 2.0 above 80 km, 2.4 between 30 and 80 km, and 2.8 below 30 km.

Our model supports the idea that vertical and meridional winds are responsible for driving seasonal changes in aerosol size and number density in Titan’s atmosphere. The aerosol fluctuations cause the seasonal cycle of the extended haze layer and albedo observed on Titan. We also reproduce a seasonal N/S albedo asymmetry, however, our model has a substantial phase shift of about one season, which may be due to the incorrect comparison of albedo in a plane parallel atmosphere to the geometric albedo. Our model reproduces the wavelength dependence of the North-South albedo asymmetry similar to the observations from Hubble, Voyager, and Pioneer [Lorenz et al., 1999]. Similarly, the wavelength dependence of the geometric albedo in our model is similar to observations, especially in the UV and visible.

Our unforced model was unable to reproduce the 20 K meridional temperature gradients and 200 m/s zonal winds observed on Titan [Achterberg et al., 2010]. This is most likely due to the CAM finite volume scheme being too diffusive at low resolution. It is also possible that it incorrectly treats angular momentum [Lebonnois et al., 2012b]. To better simulate the dynamics we added an equatorial zonal torque which accelerated the winds to 150 m/s. Despite this torque being latitudinally symmetric and constant in time, our model

reproduces a seasonally oscillating zonal jet consistent with observations. The addition of this torque also created large meridional temperature gradients consistent with observations, giving evidence of the interdependence between the temperature gradients and the winds.

Chapter 3

Titan's detached haze

3.1 Introduction

Titan's detached haze layer was first observed by Voyager as a region of enhanced scattering at an altitude of about 350 km overlying a region with a relative minimum of scattering near 320 km at all latitudes south of 45° N [Rages and Pollack, 1983]. We refer to the region of relative minimum in extinction as a gap separating the detached haze layer from the main haze layer on Titan. As discussed below the detached haze layer is something of an illusion. There is no well-defined top to the haze layer. Instead the extinction declines quasi-exponentially with altitude until the signal is lost in the noise of the observations. Likely the haze extends up to 1000 km above the surface where Cassini has directly observed high molecular weight compounds during its close flybys [Waite et al., 2007]. The haze may simply distribute itself in inverse proportion to the particle fall velocity. The gap on the other hand, is a minimum in extinction, observed above the noise level, where interesting physics may occur. Rages and Pollack [1983] retrieved an aerosol number density of 0.2 particles cm^{-3} and a particle radius of 0.3-0.4 μm in the detached haze layer between 300 and 350 km.

Cassini provided an abundance of observations of Titan's detached haze and its associated gap, the altitude of which was observed to change over time [West et al., 2011]. The gap and associated detached layer drop about 200 km in altitude during the Titan spring season between $L_s=300$ and $L_s=30$. L_s stands for solar longitude and breaks the Titan year

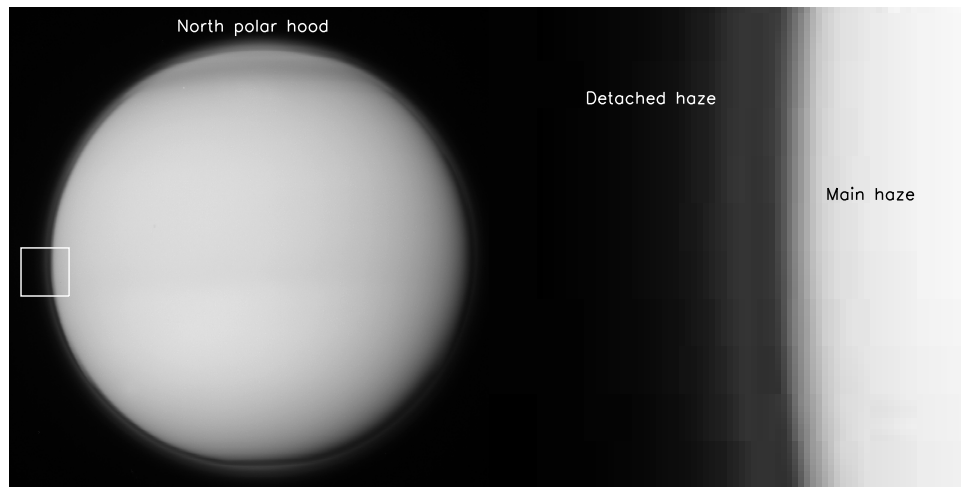


Figure 3.1: Image of Titan taken by the Cassini Imaging Science Subsystem at 445 nm on April 19th, 2011. On the left is the whole disk of Titan. A white box at the equatorial limb has been enlarged on the right to show the main and detached haze.

up into 360 degrees. $L_s=0$ indicates the northern spring equinox, which Titan went through in August of 2009. The Huygens probe landed on Titan in January of 2005.

The detached haze layer on Titan extends over most latitudes and connects to the top of the winter polar hood at 55° N latitude as seen in Fig. 3.1. The polar hood on Titan is a region of enhanced extinction associated with the downwelling branch of the pole-to-pole Hadley cell. Rages and Pollack [1983] observed a polar hood in the northern hemisphere, but not in the southern hemisphere, just after spring equinox. When Cassini arrived at Titan in the northern winter, it also observed a northern polar hood. West et al. (2013 DPS) reported that 1000 terrestrial days, or 1/10 Titan year, after the northern spring equinox, there is a polar hood in both hemispheres.

The decrease in altitude of the detached haze layer and the appearance of the southern polar hood following the northern spring equinox are effects of the more general seasonal cycle of the dynamics in which a pole-to-pole Hadley cell reverses direction with the seasons.

Evidence of the reversal of the Hadley cell after the equinox can be seen in the abundance of trace gas species [Teanby et al., 2012].

There are two types of explanations for the origin of Titan’s detached haze: microphysical and dynamical. The dynamical explanation, originally suggested using a 1-D model by Toon et al. [1992], is that the gap separating the extended haze layer from the lower main haze layer is caused by rising motions which are fast enough to suspend the haze particles. In two dimensions, upwelling in Titan’s summer hemisphere lifts haze particles low in the atmosphere, where the fall velocities are smaller than the wind speed. Above this region, where fall velocities are faster, the particles fall down into the detached layer. The aerosols are transported horizontally to the winter pole where they descend into the polar hood. Rannou et al. [2002, 2004] demonstrated this idea using a 2D GCM with coupled aerosol microphysics. Their model created a detached haze layer with extinctions close to observations. They were also able to model the seasonal cycle of the detached layer and its altitude. Although their detached haze is about 70 km below that of more recent observations leading up to the equinox, their model shows a strong drop in altitude of the haze layer around northern spring equinox, similar to that found in observations by West et al. [2011]. Cours et al. [2011] also argue for a dynamical origin of Titan’s detached haze layer based on evidence from a bimodal particle population in the detached haze. They find that the two populations of particles making up the detached haze include small particles being formed in upper atmosphere and falling into the detached layer and larger particles dynamically transported up from the main haze layer below. They conclude from analysis of scattered light in Cassini ISS and UVIS observations that the particles in the detached haze have an effective radius of $0.15\ \mu\text{m}$ around 500 km. We hypothesize that the critical test for Cours et al.’s dynamical theory is having mixed particle sizes and the near balance between vertical wind speeds and the particle fall velocities in the region of the detached haze layer, and their change with seasons as this layer changes altitude. Other expectations of the dynamical model, which were noted by West et al. [2011], include the particles following streamlines

that tend toward vertical near the summer pole. Another is that the particles fall out of the haze layer during the change in the meridional circulation around equinox. These have not been observed.

Table 3.1: Aerosol properties in the detached haze. The mass density of aerosols are derived assuming a particle density of 0.8 g/cm^3 and that the radii given are spherical or equivalent mass spherical radii. The low and mid charge cases have a charge to radius ratio of 5 and 10 electrons per micron, respectively. The 350 km values from Lavvas et al. [2009] are below the detached haze layer in their model.

Study	Season of observation	Altitude (km)	Effective radius (nm)	Number density $\#/\text{cm}^3$	Mass density g/cm^3
Rages and Pollack [1983]	Spring (Ls=18)	350	300	0.2	1.8×10^{-14}
Lavvas et al. [2009]	Winter (Ls=300)	520	40	30	6.4×10^{-15}
Lavvas et al. [2009]	Winter (Ls=300)	350	120 (spherical) 450 (fractal)	20	2.3×10^{-13}
Cours et al. [2011]	Winter - Spring (Ls=305-356)	514	150		
Aggregate inputs	Winter (Ls=300)	350	663	0.2	1.2×10^{-13}
Low charge	Winter (Ls=300)	350	423	1.9	1.8×10^{-13}
Mid charge	Winter (Ls=300)	350	255	6.5	1.9×10^{-13}
Best fit [Larson et al., submitted]	Winter (Ls=300)	350	210	5.8	8.2×10^{-14}

Lavvas et al. [2009] suggested an alternative explanation for the detached haze layer based on microphysics. They used a one-dimensional microphysical model to show that coagulation of spherical monomer particles into fractal aggregates can lead to a deficit in

extinction, since larger fractal particles, created by coagulation as particles descend, tend to have less extinction per unit mass. This model reproduces a deficit in the vertical extinction profile that is very similar to observations by Liang et al. [2007]. Their best fit model parameters include a particle size of 40 nm and a number density of 30 particles cm^{-3} in Titan’s detached haze layer. Lavvas et al. [2009] did not offer an explanation for the vertical movement of the haze. We hypothesize that the critical test for the Lavvas microphysical theory is the sudden shift from monomers to fractal particles in the region of the gap, or in other words, a change in particle shape across the gap.

Properties of the particles that form the detached haze layer retrieved from the studies mentioned and our simulations are compiled in Table 3.1.

In this paper we use a 3-dimensional GCM with coupled aerosol microphysics to extend the efforts of the previous studies explaining the origin and seasonal evolution of Titan’s detached haze layer. We can test the hypotheses for the formation of the detached haze, mentioned above, with our model by investigating the microphysical and dynamical conditions under which our model produces a detached haze layer. We compare our model results with observations to determine a formation theory of the detached haze and explain its seasonal cycle.

3.2 Description of the model

In this study we use the Titan CAM/CARMA model described in Chapter 2. Our microphysical model simulates aerosol coagulation and sedimentation. The simulated aerosols are radiatively and dynamically coupled to the GCM. Aerosol mass is input at the top of the model, around 560 km. As the aerosols move through the atmosphere they grow into aggregates with a fractal dimension of 2. When we refer to the radius of a particle in this chapter, we are referring to the equivalent mass spherical (ems) radius of these aggregate particles. The aerosols in Titan’s atmosphere are thought to be electrically charged, which inhibits coagulation [Borucki et al., 1987]. The coagulation rate of the aerosols in our model

is greatly affected by the ratio of electrical charge placed on the aerosols to their radius.

Table 3.2: Columns 1-3 provide microphysical parameters of our simulations. Columns 4 and 5 define the particle radius at 350 km and whether the simulation produces a detached haze layer. The effective radius given is the particle area weighted average equivalent mass spherical radius.

Simulation	Charge (e^- / 1000nm)	Prod. rate ($\times 10^{-14}$ g $\text{cm}^{-2} \text{ s}^{-1}$)	Input size (nm)	R_e (nm) at 350 km	Detached haze pres- ence
Base case	15	3	42	208	weak
Best fit	7.5	1	42	210	none
Aggregate input	15	3	663	663	strong
Molecular input	15	3	2	213	none
Low Mass	15	1	42	137	weak
High Mass	15	5	42	284	weak
Low charge	5	3	42	423	strong
Mid charge	10	3	42	264	strong
High charge	20	3	42	170	none

We run several simulations of the extended haze exploring the impact on the haze of three microphysical parameters; the charge to particle radius ratio, the aerosol mass production rate, and the particle size input at the top of the model. Our base case simulation uses a charge to radius ratio of $15 e^-/1000\text{nm}$, a mass production rate of $3 \times 10^{-14} \text{ g cm}^{-2} \text{ s}^{-1}$, and a starting size of 42nm particles. From this base case, we run sensitivity tests varying one parameter at a time. We also analyze the best fit case from Chapter 2. Table 3.2 describes the simulation microphysical parameters, the effective radius of the particles at 350 km, and whether they produce an equatorial detached haze layer during the season of the Huygens landing (northern winter, $L_s=300$). The effective radius, R_e , is the area weighted mean radius of equivalent mass spherical particles.

We consider a simulation to have produced a detached haze if there is a level of greater extinction than the level below it at 525 nm, i.e., a positive slope in the vertical extinction profile. If this positive slope is only one model level thick, we consider it a weak detached

haze layer. A simulation has a strong detached haze layer if the positive slope in the vertical extinction is at least two model levels thick. Another key test of this model is the ability to produce a summer polar detached haze layer at the observed altitude.

3.3 Results

3.3.1 Haze vertical profiles

To test the detached haze formation theories described above we run a series of simulations with different microphysical parameters. We expect that if the detached layer is due to changes in the size and shape of the particles as they coagulate, then simulations with the greatest particle size changes should produce the greatest gap in the vertical extinction profile. However, if the formation of the detached haze is controlled by dynamics through the balance of vertical winds and particle fall speeds, then we predict that the detached haze and gap will be independent of changes in particle size and shape. Furthermore, this balance in velocities will be found in the region of the detached haze in the model.

Many, but not all, of our simulations produce a detached haze layer as seen in their equatorial vertical extinction profiles. Fig. 3.2 shows the extinction profiles from our simulations compared to Voyager and Cassini observations during the northern winter ($L_s=300$). For purposes of simplifying the figure, we have scaled the Cassini extinction profiles by a factor of two to account for the wavelength difference between Cassini observations and Voyager observations. We show our simulations at a similar wavelength to the Voyager data. The top panel in Fig. 3.2 compares simulations with varying input sizes at the top of the atmosphere. The middle panel compares simulations with varying aerosol mass production rate and the bottom panel compare simulations with a varying charge to radius ratio.

Only some of our simulations produce the local minimum in extinction, a gap, defining the base of the detached haze layer. The presence of the gap is informative about the nature of the detached haze because it allows us to distinguish between the theories of detached haze

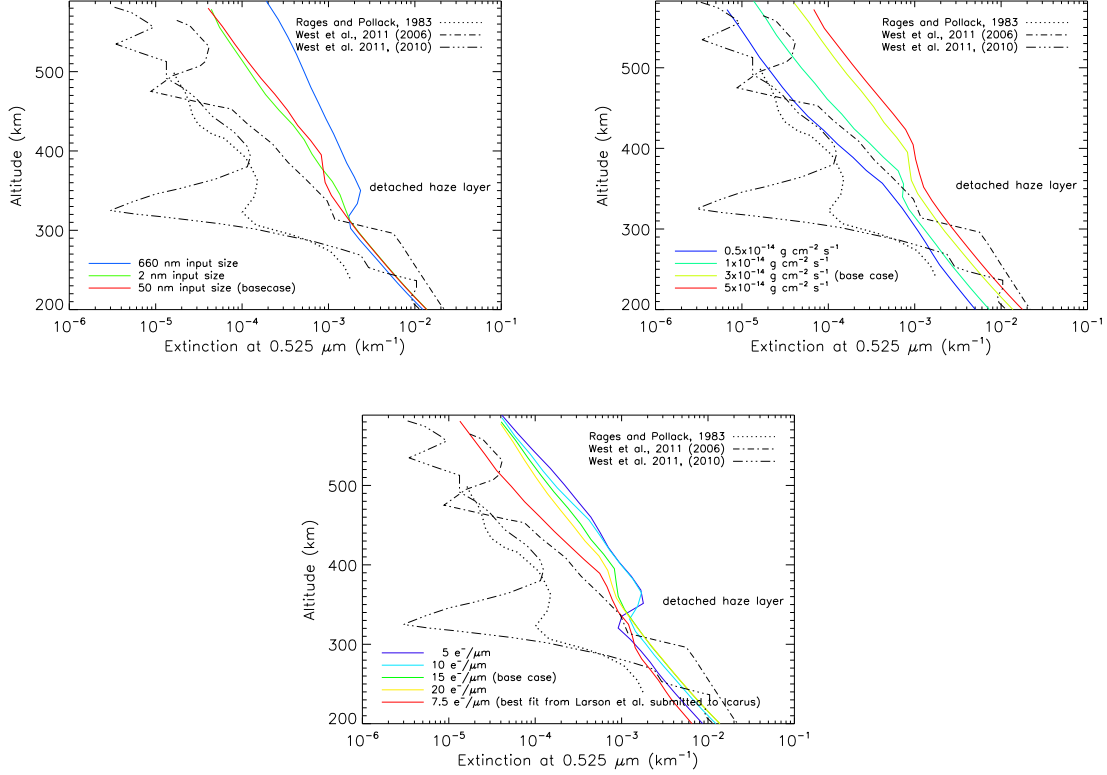


Figure 3.2: Aerosol extinction at 525 nm in Titan’s stratosphere compared with observations from Voyager [Rages and Pollack, 1983] and Cassini [West et al., 2011]. The Cassini data, obtained at a wavelength of 340 nm, have been multiplied by a factor of 2 to account for the difference in wavelength. The factor of 2 is consistent with the wavelength dependence of the extinction in the simulations.

formation. The altitude of the gap in our haze simulations, where it exists, is consistent with that seen by Voyager and the 2006 Cassini observation (Fig. 3.2). However, our simulated gaps in the haze are well below the altitude of the local minimum in extinction seen by Cassini in 2006.

Observations of Titan’s vertical extinction profile are more variable than our simulations. The 2006 vertical extinction profile provided by West et al. [2011] show Titan’s detached haze layer as an overabundance compared with the exponential line. The gap in

the haze is a clear deficit from this exponential line. The 2010 image analyzed by West et al. [2011] shows a detached haze layer that is only a local maximum because of the deficit (gap) below it. The local maximum in this image lies on the exponential line. Our simulations have a different morphology compared to the West et al. [2011] observations. Our simulations that produce a detached haze layer appear to have a step function in the haze extinction. The exponentials that would fit our simulations above and below the gap would have a similar slope but different intercepts. This is shown clearly when we plot the mass density over air density for the aggregate input simulation in Fig. 3.5. There are two relatively constant lines, one above and one below the haze gap.

The simulated local minimum extinction in the haze is not as small as Cassini observations. The vertical resolution in our model is quite low at the altitude of the gap, and large discontinuities get smoothed by the model. The local extinction minimum in our model is generally resolved by only two model layers. This resolution is not sufficient to support a minimum that deviates from the average by much more than a factor of 2.

There are three simulations that produce a very clear gap in the haze. These are noted in Table 3.2. They include the simulation in which we input $0.67 \mu\text{m}$ equivalent mass sphere aggregate particles at the top of the model and the simulations with charge to radius ratios of 5 and 10 electrons per micron. The common theme between these three simulations is attaining a large particle size near 350 km. A low charge on the particles allows for growth from coagulation. These simulations have a mass production rate, $3 \times 10^{-14} \text{ g cm}^{-2} \text{ s}^{-1}$, which is a factor of three larger than the best guess aerosol production rate from Chapter 2.

Interestingly, the best guess simulation, which produces the best agreement with data largely from lower altitudes, does not produce a strong gap in the haze despite the low charge to radius ratio of 7.5 electrons per micron. This charge to radius ratio is between two values that clearly produce a gap in the haze. The lower mass production rate in the best guess simulation, $1 \times 10^{-14} \text{ g cm}^{-2} \text{ s}^{-1}$, is a factor of three lower than the low and mid charge simulations. Our model suggests that higher production rates are necessary to increase the

particle size and create an observable detached haze layer. We tested the impact of mass production rate (Fig. 3.2, middle panel) and found that with a charge to radius ratio of 15 electrons per micron, even large mass inputs show no local minimum or maximum in the haze. This is evidence that large a particle size is important in producing a gap in the haze. Of the tested parameters, particle charging seems to have the largest effect on our model's ability to reproduce a detached haze because it greatly affects the aerosol size distribution.

We earlier suggested that the Lavvas theory of the detached haze required that fractal formation occurred in the region of the gap, while the idea of Cours required that small particles from high altitudes mixed with larger ones below. While we cannot exclude these possibilities on Titan, they are certainly not required as shown in our model for the aggregate case in Table 3.1 and Fig. 3.2. The aggregate case produces a strong detached haze layer as shown in Fig. 3.2. However, in this case the particle size is independent of altitude as shown in Fig. 3.3, panel c. Therefore there are not smaller particles aloft nor fractals forming in the gap, and the theories of Lavvas and Cours do not apply to this model simulation.

Generally, Fig. 3.3 shows that the extinction profile tracks the aerosol mass density profile. The mass densities (solid lines) show a local minimum, creating a detached layer, in most simulations. For cases in which the effective radius decreases below 200 nm with increasing altitude, the extinction decreases more rapidly with increasing altitude than does the mass because the optical efficiency (extinction per unit mass) of the particles decreases as the particle radius drops toward the Rayleigh limit. In some cases the change in effective radius with altitude can mask the impact of the mass density on the extinction and prevent a clear detached layer from being visible. For example, the two simulations with the lowest and highest charge to radius ratio (Fig. 3.3, d and f), produce very different detached haze layers. The low charge case produces a strong detached haze layer and the high charge case produces essentially no detached haze layer as seen in the extinction profiles. In the low charge simulation, the particles in the detached haze layer (350 km) have a unimodal distribution and are about 200 nm in radius. In the gap below (320 km), the particles have

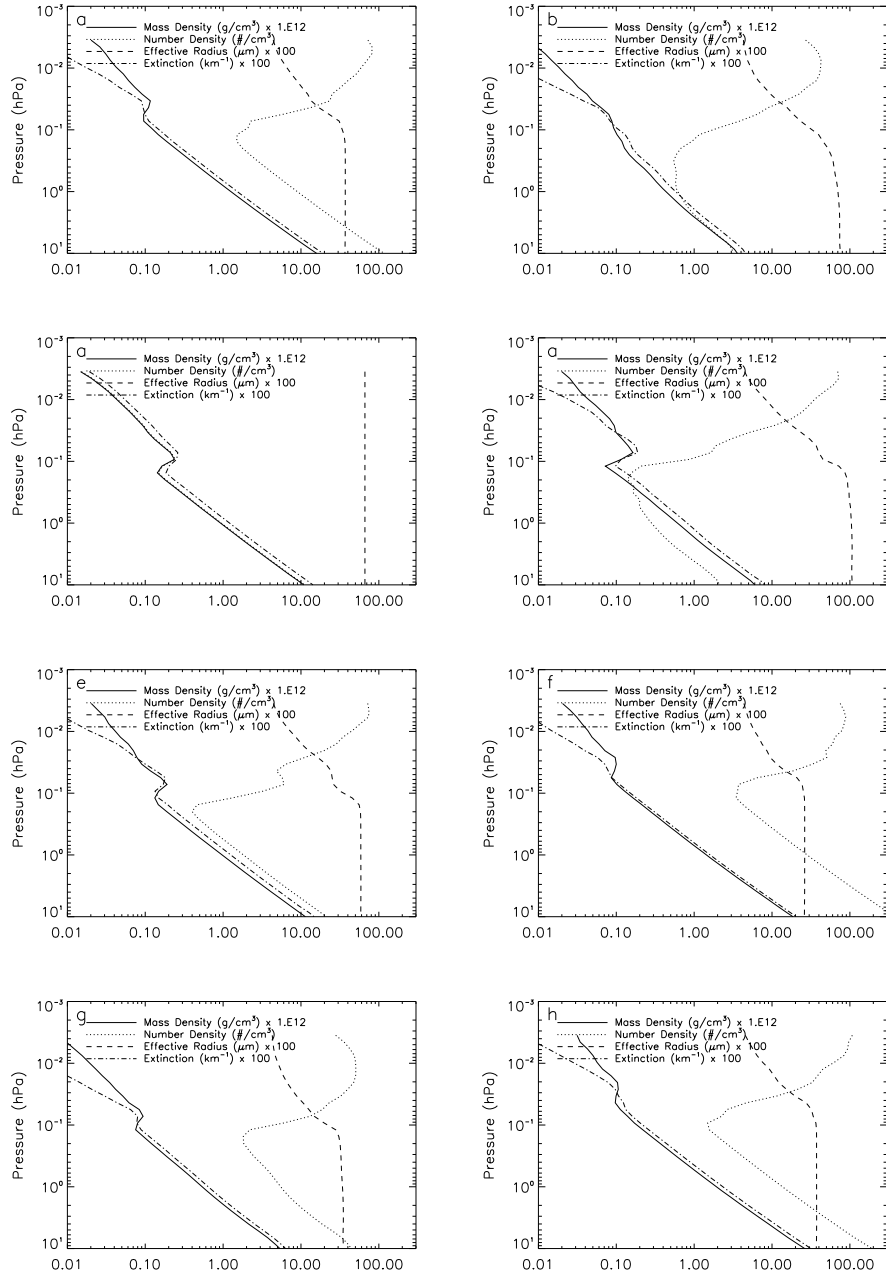


Figure 3.3: Comparing the vertical profiles of the mass density, equivalent mass spherical effective radius, number density, and extinction between model simulations. The simulations from left to right and top to bottom are: base case, best fit, aggregate input, low charge, mid charge, high charge, low mass, and high mass.

a bimodal distribution with a significant contribution from larger micron sized particles. In

this case, the detached haze particles are in the size range observed by Rages and Pollack [1983] (see Table 3.1). Above about 200 nm radius, there is no gain in scattering efficiency at a wavelength of 500 nm as particles grow larger (see Appendix A.4), therefore the extinction responds strongly to the mass density profile. The high charge case has smaller particles, but a similar pattern, with a unimodal particle size distribution of 70 nm radius in the detached haze (385 km) and a bimodal distribution in the associated gap (350 km) with the larger particles up to 300 nm. Particles in this size range have a significant difference in scattering efficiency, which tends to obscure the response of the extinction to the changing mass density in our simulations.

Fig. 3.4 shows the size distribution at altitudes spanning the detached haze layer during the northern winter for the low charge and high charge simulations. The shift to a bimodal distribution in the gap in the haze is apparent in both simulations. This bimodal distribution is consistent with observations at 500 km from Cours et al. [2011], and are suggestive of upwelling winds. This suggests the particles in the gap come from two different populations and is consistent with convergence in vertical transport of the particles. For reasons suggested below our model does not produce a detached haze layer at 500 km as observed by West et al. [2011] and Cours et al. [2011]. However one would expect both varying particle size, and varying mass density to be factors in such high altitude haze layers.

3.3.2 Analytical description

In order to clarify the causes for the detached haze layer we construct a simple analytical description of the aerosol number density in the region of the detached haze and gap. For simplicity, let us consider the aggregate input case in which the particle sizes are fixed, so the number density, mass density, and extinction are all proportional (Fig. 3.3, c). The aerosol extinction in much of Titan's upper atmosphere is observed to be fit by an exponential [West et al., 2011]. The air density is also an exponential with altitude. In Fig. 3.5 we plot the

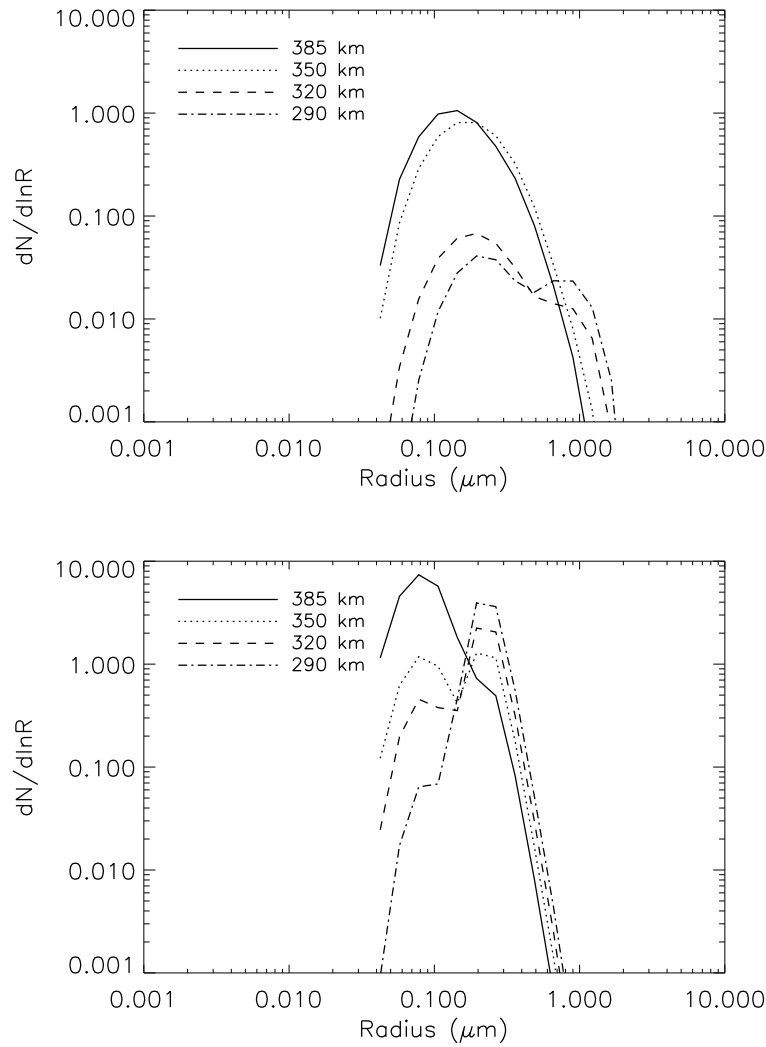


Figure 3.4: Size distribution of low charge simulation (top) and high charge simulation (bottom) spanning the altitude of the detached haze layer.

aerosol mass density over air density along with the vertical velocities in the aggregate input simulation. The left panel is the northern winter (the same time as the simulations in Fig. 3.3), and the right panel is during the northern spring. The mass density / air density (dash dot line) is roughly constant above and below the detached haze and gap, showing that

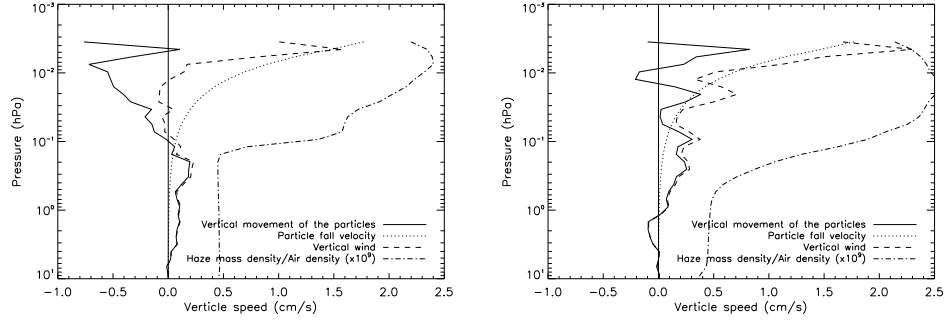


Figure 3.5: Simulated particle mass density/air density (dash dot line), vertical wind (dashed line), fall velocity (dotted line), and vertical particle movement (solid line) during the northern winter (left) and spring (right) from the aggregate input simulation.

the mass (and extinction) is proportional to the air density. The particle velocity (dotted line) is inversely proportional to the air density. The vertical winds (dashed line) however, increase with altitude. Where these two are equal the particles are suspended and we get convergence, upwelling below and particles falling above. The net vertical movement of the particles (solid line) is the vertical wind velocity minus the particle fall speed and looks somewhat like a Gaussian in the location of the gap.

We construct a simple 1D model to illustrate this gap formation. The steady state aerosol continuity equation in the absence of all processes except vertical transport by winds and gravity is given by Eq. 1. The solution to this equation is that number density, N , is proportional to the inverse of total velocity. The constant of proportionality is the product of the number density, N_{top} , and velocity, V_{top} , at the top of the model. Assuming the vertical wind speed is small compared with the fall velocities at the top of the model, V_{top} is just our fall velocity at the top of the model, which is 2 cm/s. We can use the exponential fits to each observation to get the value of N_{top} .

$$\delta(NV)/\delta z = 0 \quad (3.1)$$

However, the particle fall velocity is approximately equal to a constant divided by the air density, $V_{top}\rho_{top}/\rho$. Therefore, assuming the wind speed is small compared with the fall speed,

$$N = V_{top}N_{top}/(V) \quad \sim N_{top}\rho/\rho_{top} \quad (3.2)$$

Where the vertical wind is negligible, the number density (and extinction if radius is constant) will vary in proportion to air density as shown to be the case in Fig. 3.2 and 3.5. The vertical particle movement is approximately constant everywhere except the location of the gap. At this location, we model the vertical wind as a Gaussian, as suggested by the model winds illustrated in Fig. 3.5.

$$V = V_{top}\rho_{top}/\rho + W \quad (3.3)$$

Here W , is a Gaussian function peaked at the altitude of the gap. We fit this simple model to the extinction profiles from West et al. [2011] in Fig. 3.6. The left panel is from the 2010 observation and the right panel is from the 2006 observations. Our simple model with a Gaussian wind profile reproduces the gap in the haze and its associated detached layer.

Table 3.3 gives the parameters for the analytic model that fits the extinction profiles from these two observation. To compute the extinction ($Ext = Q_{scat}\pi r^2 N$) we assumed that the scattering efficiency was 11.5 (See Appendix A.4) and the particle size was 663 nm, the same as the aggregate input case. These values make the extinction 0.013 times the number density. The altitude of the gap is the altitude where the Gaussian profile peaks, which in our model relates to where the particle fall speed is matched by the vertical wind speed. The depth of the extinction minimum in the gap is controlled by the ratio of V_{top}/W_0 , i.e., the magnitude of the background wind compared with the Gaussian wind maximum. The

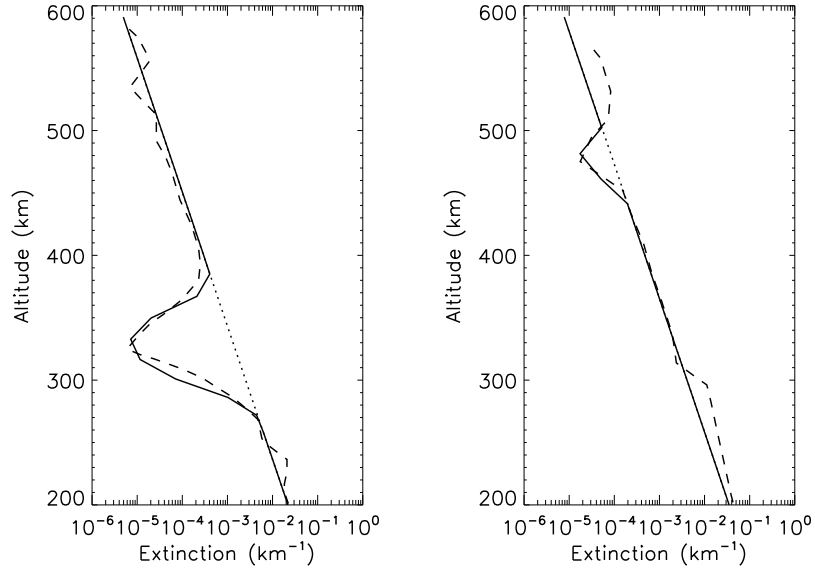


Figure 3.6: Fitting our simple analytic model to the haze extinction profiles of West et al. [2011]. The extinction profiles are the dashed lines. The dotted lines are the exponential fits to the observations. The solid line is our simple model fit assuming a Gaussian wind profile with parameters given in Table 3.3 and that the wind speed is small compared with our fall velocities.

thickness of the gap is controlled by σ , the variance of the Gaussian profile. Outside of the region of the gap, the Gaussian wind, W , is negligibly small.

The exponential fits to the extinction profiles given by West et al. [2011] have the same slope. This indicates that the decrease in density with altitude, i.e., the scale height, is consistent between the observations. Thus the stratospheric equatorial temperatures must not be changing greatly during this time interval. The exponential fits have different intercepts, however, indicating that either the number density or vertical wind has increased between the observations. The number density could change due to changes in the photochemical production rate as Titan moved through equinox or changes in horizontal advection of par-

Table 3.3: The parameters of our analytic model that fit the 2006 and 2010 extinction retrievals by West et al. [2011].

Parameter	2010 Observa- tion	2006 Observa- tion
N_{top} (cm^{-3})	1.9×10^{-5}	3.0×10^{-5}
V_{top} (cm/s)	2	2
W_0 (cm/s)	1.4	1.4
σ (km)	19	9
z_0 (km)	330	474

ticles. Note that the ratio of the Gaussian wind magnitude to V_{top} that best fits the gap in the haze is consistent between the two observations, even though the gap in the 2010 observation is over an order of magnitude deeper. This indicates a similar balance between the vertical upwelling wind speed and fall velocities at those two altitudes. Since the fall velocities are much higher aloft, the upwelling must also be higher at 470 km in the 2010 observation than it was at 330 km in the 2006 observation.

In the observations of West et al. [2011], the exponential fits to the extinction above and below the gap are identical (Fig. 3.6). However, in our simulations there is often an offset (Fig. 3.3). The offset cannot be reproduced by our simple analytic model. We can apply our analysis from above to our model profiles. The increase in the offset between the upper exponential slope and the lower exponential slope indicates that there are more particles above. Thus we also have a loss term in our simulations due to horizontal advection that transports particles towards the winter polar hood. This loss occurs where the particles are suspended, allowing horizontal transport to dominate the loss. The presence of the offset in our simulations is not observed by West et al. [2011], indicating that our GCM is removing too many particles due to advection toward the poles compared with observations. The horizontal wind speeds on Titan must be lower than in our simulations.

3.3.3 Seasonal cycle

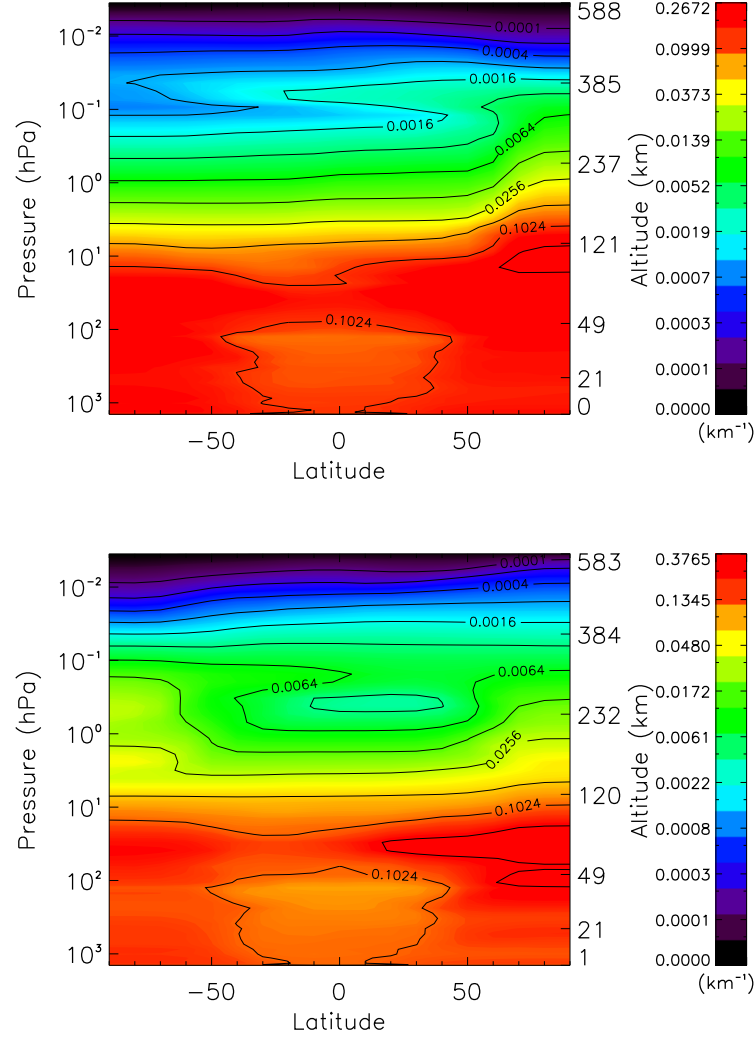


Figure 3.7: The zonally averaged extinction from the low charge simulation during the northern winter (top) and spring (bottom). The bottom panel is at 1000 days after equinox which corresponds to the time of the West et al. (2013 DPS) observation.

Fig. 3.7 shows the 2D contour of extinction in our aggregate input simulation at two different times. This simulation produced a strong gap and associated detached haze layer

and northern polar hood during the northern winter (LS 300) near 350 km. In the northern (winter) polar region, the detached haze merges with the polar hood, which is a region of enhanced aerosol extinction that extends up above the main haze layer. The northern polar hood is a region of strong downwelling in our simulations of Titan's atmosphere. The detached haze layer extends from the southern pole northward. However it is noteworthy that the detached haze does not extend to the southern pole in all microphysical simulations. Figure 3.8 shows the mass density of the haze in our two simulations with the most pronounced detached haze layer and associated gap. These are the aggregate input case (left) and the low charge case (right). The asterisks in the figures display a local maximum in the haze density, i.e., the detached haze layer. The detached layer in our extends from the edge of the polar hood (60° N) south. In the aggregate input simulation, the detached haze layer stops at 50° S. In the low charge simulation the detached haze layer extends all the way to the southern pole, which is consistent with observations from Cassini between 2005 and 2012. That many of our simulations do not produce a detached haze that extends all the way to the pole likely reflects two problems with the model. The vertical resolution is low which makes resolving all be the strongest gaps in the haze difficult. Secondly, poles are places of convergence in GCMs and therefore have special filters that can further diffuse inhomogeneities such as the detached haze and gap.

The extinction in our model 1000 days after equinox (Fig. 3.7, bottom panel), which corresponds to observations from West et al. (2013 DPS), shows a more symmetrical haze pattern. A polar hood has formed in the southern hemisphere, and the northern polar hood is present, but dispersing. We see a local minimum, or gap, in the vertical extinction profile at all latitudes between the polar hoods at an altitude of about 270 km. The local maximum of the associated detached haze layer is near 300 km. Our model reproduces the detached haze and polar hood distribution seen by West et al. (2013 DPS), who reported a polar hood present in both hemispheres 1000 days after the equinox. The formation and dispersion of the polar hoods is associated with strong upwelling and downwelling at the poles due to

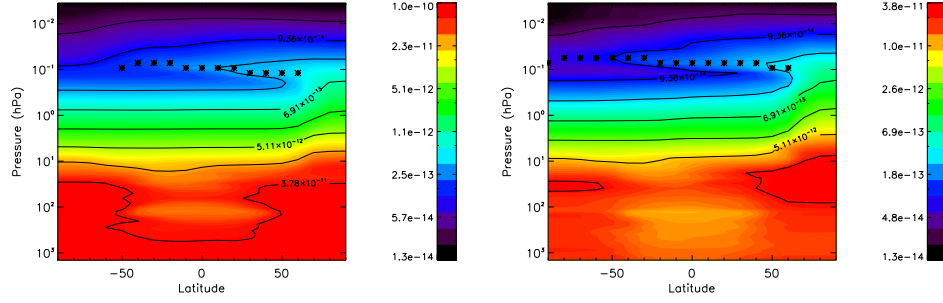


Figure 3.8: Mass density plots of our aggregate input (left) and low charge (right) simulations during the northern winter. The asterisk symbols show the local maximum in the vertical mass density profile.

the overturning pole-to-pole Hadley cell. Our model reproduces these features, which can be seen in the vertical and meridional wind profiles in Fig. 3.9 and in the meridional mass stream function averaged over Titan's seasons in Fig. 3.10.

Titan's equatorial detached haze layer was observed by Cassini to decrease in altitude as Titan moved through equinox [West et al., 2011]. Fig. 3.11 plots the altitude of the simulated detached haze over one Titan year for the aggregate input and low charge cases. Here we use the same definition of the detached haze layer as West et al. [2011] and plot the altitude of the local maximum of extinction. These simulations produce an extended haze that decreases in altitude, until merging into the main haze layer 4-5 earth years after equinox (LS 54). In the simulations the detached haze layer reforms at its original location high in the atmosphere soon after merging with the main haze, about 5-5.5 years after equinox (Ls=60-66). We predict Titan should reform its detached haze layer between mid 2014 to early 2015. The pattern of these events is consistent for each of the simulations listed in Table 3.2 that produced an extended haze layer, however the exact timing and vertical movement varies somewhat based on the simulation. Part of this variation is due to the visibility of the detached haze layers described above. Another source of variability

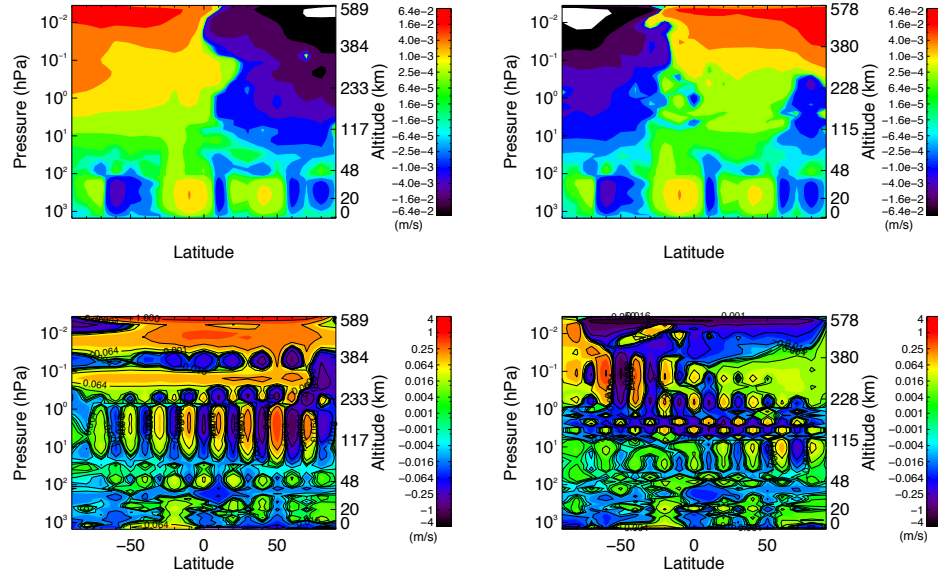


Figure 3.9: Zonally averaged vertical winds (top) and meridional winds (bottom) from the aggregate input simulation during the northern winter (left) and spring, 1000 days after equinox (right). These winds were averaged over 1/50th of a Titan year.

is the low vertical resolution and low model top in our simulations which makes weak gaps at higher altitudes, such as the gap observed by West et al. [2011] in 2006, difficult for our model to resolve.

The response of the haze to the vertical winds can be seen in Fig. 3.12, which plots the time evolution of the equatorial haze mass density (g/cm^3) in the top panel and the equatorial vertical particle movement (cm/s) in the bottom panel for the aggregate input simulation. The vertical particle movement is the combination of fall velocity and vertical wind which shows the true movement of the particles. Overlain on these plots are the data from Fig. 3.11 indicating the location of the detached haze layer. The top panel in Fig. 3.12 clearly shows the equatorial detached haze layer as a local maximum in the vertical mass

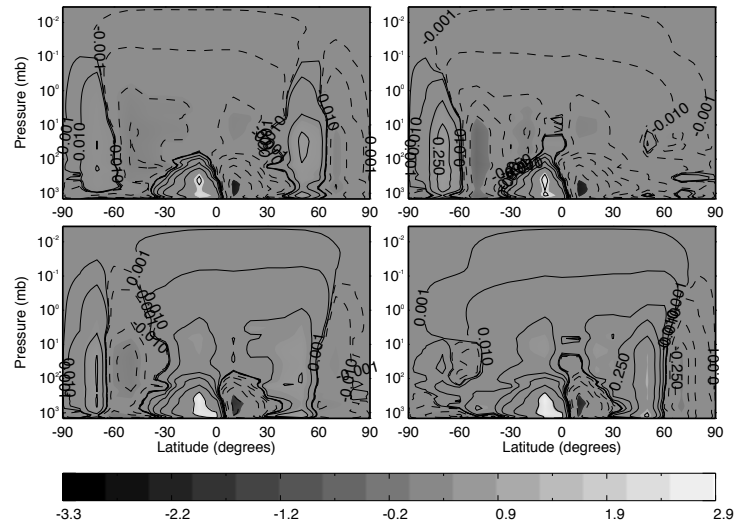


Figure 3.10: Meridional mass stream function ($\times 10^9$ kg/s) averaged over one Titan season during the spring, summer, fall, and winter (clockwise starting from the upper left panel).

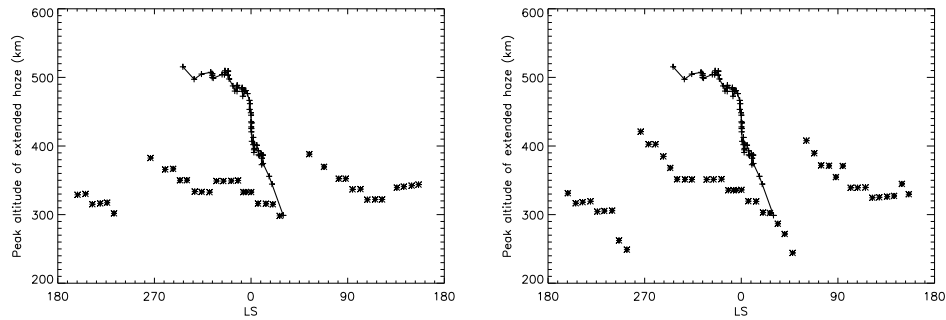


Figure 3.11: The seasonal cycle in the altitude of the equatorial detached haze layer in our aggregate input (left) and low charge (right) simulations. Equinox is at LS 0 in the center of the figure. The asterisk symbols are our model simulations, while the plus symbols connected by a line are observations from West et al. [2011].

density profile. The vertical movement of the particles is dominated by the fall velocities

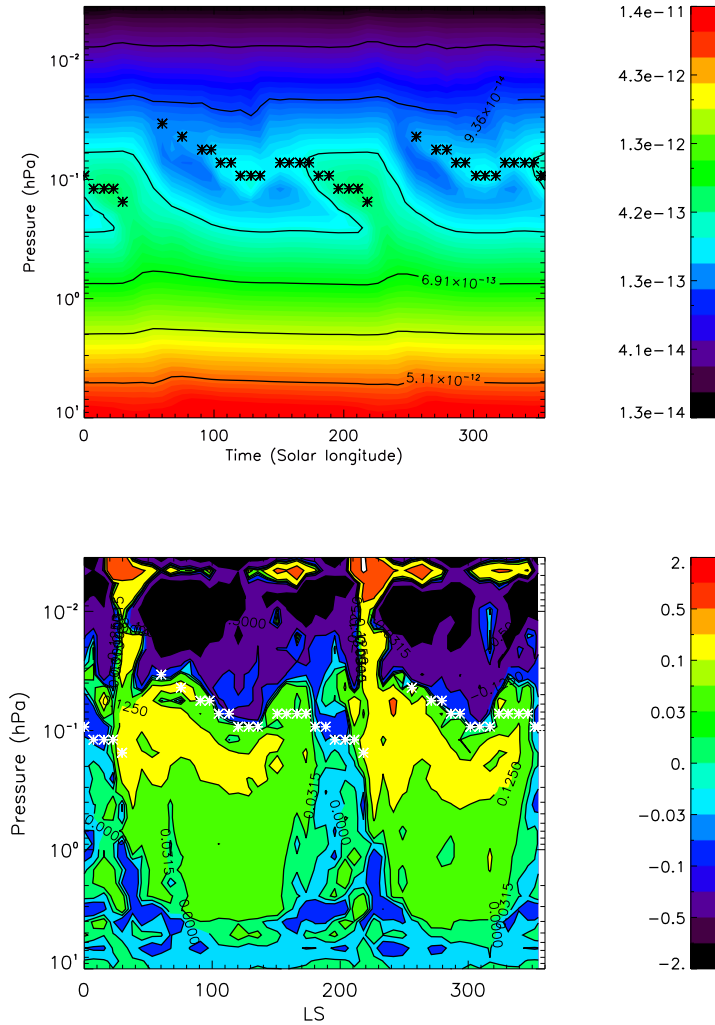


Figure 3.12: Time evolution at the equator in our low charge simulation over one Titan year. Top panel shows the vertical variation in equatorial aerosol mass density (g/cm^3). The bottom panel shows the vertical particle movement (cm/s) (vertical wind + fall velocity). The asterisk symbols map the altitude of the simulated equatorial detached haze layer from Fig. 3.11.

above the detached haze and the vertical upwelling winds below the detached haze layer (Fig. 3.12, bottom panel). Upward winds cause convergence of the particles when they are

similar to fall speeds, so the vertical particle movement is close to zero, creating the detached haze layer. Just before equinox ($L_s=0, 180$) in our simulations, the upwelling winds diminish and the particles fall for about three Earth years, eventually blending with the main haze layer. After this ($L_s \approx 30, 220$), there is such strong upwelling in the tropics that the vertical movement is upward at all levels and a detached layer does not occur. This strong upwelling slowly subsides and the balance between upwelling wind and the particle fall velocities is reestablished again at high altitudes.

Even though there is upwelling in the summer hemisphere and downwelling in the winter hemisphere due to the global Hadley cell (Fig. 3.10), we observe little latitudinal dependence in the altitude of our detached haze. Our low charge simulation which has a detached layer that extends all the way to the pole has a maximum equator to pole difference in altitude of 17 km. This is consistent with the 7-32 km equator to pole altitude difference from West et al. [2011]. The consistent altitude of the detached haze is likely due to the relatively strong meridional winds that transport haze to the winter pole (Fig. 3.9). The streamlines (Fig. 3.10) in our simulations are relatively horizontal between the midlatitudes. For example, a 1 m/s meridional wind would transport material from the equator to the edge of the polar hood in about 30 earth days. In contrast, the vertical wind outside the polar hood is generally less than 0.5 cm/s at the altitude of the detached haze (Fig. 3.9). Similarly, the fall velocities are a small fraction of a centimeter per second. A net 0.6 cm/s downward vertical movement over 30 days as the particles moved from the equator to the winter pole would drop the detached haze and gap by 16 km, which is the altitude difference of the detached haze in our model between the equator and edge of the polar hood.

3.4 Discussion

From this analysis we can construct a formation theory for the detached haze layer in Titan's atmosphere based on the balance of vertical upwelling wind and particle fall velocities, similar to that given by previous studies [Toon et al., 1992, Rannou et al., 2002,

Cours et al., 2011]. The detached haze layer is really defined by an extinction minimum or gap within a column of haze (West et al. 2013 DPS). The layer is fed from the top by chemical production of aerosols above the top of our model (560 km), and from the bottom by transport from the main haze layer if there are upward motions. There is some level at which the vertical wind speed equals the fall speed of the particles. Above this level particles cannot be transported upward, they only fall from above because they have higher speeds at higher altitudes, and you get a maximum. There are secondary processes at work on Titan, not covered in our simple analytic model, but included in our numerical simulations. If the particles are changing size in the 100 nm size range in the region of interest, then the gap may not be observed in the extinction profile. This is because the increase in extinction due to increasing scattering efficiency as the particle grow larger as they fall downward can cancel out the decline in extinction due to the mass density reduction as the particles are suspended by the winds. However, fractal particles larger than a couple hundred nanometers do not change their scattering efficiency. The particles might either be large because they grew above the top of the model, or because coagulation is relatively fast, due to low charge and a large amount of mass input. The altitude of the gap falls, as the upwelling wind speeds decline toward equinox. As wind speeds drop the gap drops, because for the same size particles the balance will occur lower in the atmosphere. The altitude and seasonal cycle of the detached haze layer and associated gap below it can be explained as a balance between the particle fall velocities and the upwelling vertical winds.

The latitudinal uniformity of the detached haze layer is due to the slow vertical motions compared with the meridional motions in Titan's atmosphere associated with the pole-to-pole Hadley cell. We find that the seasonal cycle of the detached haze altitude is driven by the overturning Hadley cell. Our simulations predict the detached haze layer will disappear between early 2012 and early 2014. The haze dissipated by the beginning of 2013. This is within the range of our model simulations. Our model predicts the reappearance of the detached haze layer at high altitudes between mid 2014 and early 2015. This is qualitatively

consistent with the simulations from Rannou et al. [2002], however our model predicts a later disappearance of the detached haze by a few years.

The discrepancies between our model simulations and observations can be explained within this framework. Our model produces a detached haze layer at too low of an altitude, probably because the vertical winds are too weak above about 400 km. Also, the vertical resolution of our model is quite low above 400 km making gradients difficult to generate. Furthermore, the haze gaps in our simulations do not have low enough extinction due to the low vertical resolution in our model. The temperatures at the top of the model are lower than the observations which will slightly affect the fall velocities and microphysics, which may impact the details of the detached haze distribution as well.

Chapter 4

Application to Titan's dunes

4.1 Introduction

The Cassini mission made an unexpected discovery when it found evidence of linear dune fields on Titan's surface [Lorenz et al., 2006b, Radebaugh et al., 2008]. The radar dark dunes have a mostly tropical distribution between ± 30 degrees latitude that extends to all longitudes. Cassini has mapped thousands of dunes throughout its mission, many of which are observed to divert around topography. The crestline orientations of the dunes and their interaction with topography allow scientists to estimate the dominant wind direction on the surface of Titan. There is some consensus in the community that the dune-forming winds produce a westerly (West to East) net transport [Lorenz et al., 2006b, Radebaugh et al., 2008, Barnes et al., 2008, Rubin and Hesp, 2009, Tokano, 2008, 2010].

However, there is an active debate about the dune forming wind regime. This debate has been guided by several studies of Earth dune fields considered analogous to the Titan dunes including those in Namibia, the Sahara, the Serengeti, and China [Lorenz et al., 2006b, Radebaugh et al., 2008, Rubin and Hesp, 2009]. The linear dunes have been considered analogs to Namibian dunes due to morphological similarities [Radebaugh et al., 2008]. The presence and morphology of these dunes can aid our understanding of the dune forming winds on Titan. A description of the relevant theories can be found in Rubin and Hesp [2009] and Tokano [2010]. Complicating this active debate about the surface wind regime is the fact that GCMs have historically not been able to reproduce westerly surface winds

in the tropics [Friedson et al., 2009, Lebonnois et al., 2012a, Tokano, 2008, Newman et al., 2011].

The Titan surface wind regimes that have been proposed, mostly based on studies of morphologically similar Earth dune fields, are as follows: westerly winds with a fluctuating tidal component [Lorenz et al., 2006b]; bimodal (coming from two directions) westerly [Radebaugh et al., 2008]; bimodal easterly [Tokano, 2008]; unimodal westerly [Rubin and Hesp, 2009]; bimodal and slightly westerly [Tokano, 2010]; and both unimodal and bimodal westerly winds (Radebaugh et al. 2010 LPSC abstract). Some of these theories are in conflict with maximum gross bedform-normal transport theory (MGBNT) [Rubin and Hunter, 1982, Rubin and Ikeda, 1990, Fenton et al., 2014]. Rubin and Ikeda [1990] argue that longitudinal dunes only form in a specific type of wind regime with bimodal winds having roughly equal magnitudes and divergence angles of greater than 90 degrees. It is not certain that Titan's dunes are longitudinal, but they appear morphologically similar to longitudinal dunes on Earth.

One important thing to note is that, to date, only one GCM has offered detailed results of the surface wind patterns and how they relate to the dunes [Tokano, 2008, 2010]. Tokano [2010] offers an explanation of the dune forming winds as follows. Although the net wind is easterly and usually around 1 m/s, near equinox there are strong westerly winds of slightly higher magnitude, 1.5 m/s. Although westerly winds are rare, their increased magnitude allows them to carry dune grains while the weaker easterly winds cannot, thus building the dunes. This is based on the principle that the potential saltation flux is proportional to the cube of friction velocity, thus the sand flux is very sensitive to the wind strength [Bagnold, 1941].

Other studies have offered ways to get around the seemingly difficult task of producing surface prograde winds in the tropics. Hayes et al. [2012] for example, suggest that Titan's dunes form over orbital time scales and may not be in equilibrium with the current wind regime.

In this paper we present results from a Titan CAM GCM that includes the effects of realistic topography from the latest Cassini topography map [Lorenz et al., 2013]. We show that the addition of topography can change our simulated surface wind pattern creating large areas with westerlies in the tropics.

4.2 Description of the model

We run the model at 4x5 degree resolution. See Friedson et al. [2009] for a detailed description of the model. In this study we consider simulations with two improvements to the original model, topography and an ad hoc zonal wind forcing term.

Recently a 1x1 degree topographic map was released to the community by the Cassini radar team [Lorenz et al., 2013]. This map was reduced to 4x5 degree resolution and incorporated into our model.

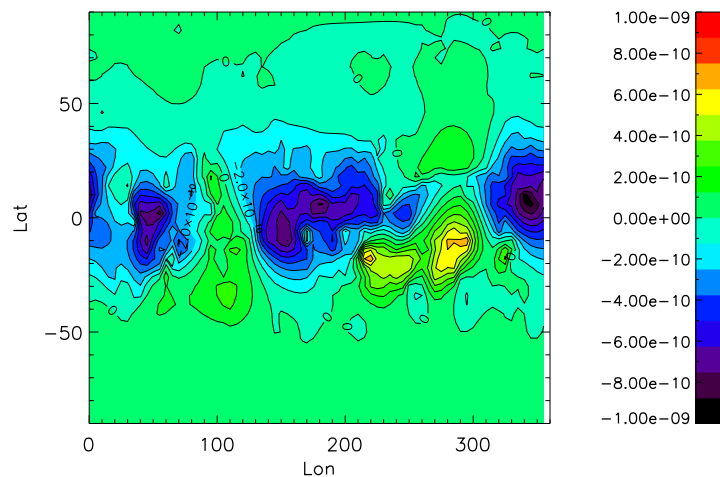


Figure 4.1: Ad hoc zonal acceleration (m/s^2) at the surface of Titan. West is positive.

Secondly, this model uses an adhoc torque to increase the magnitude of the zonal winds. This forcing was originally incorporated into the model to accelerate the stratospheric jets in order to make more realistic stratospheric aerosol simulations (see Chapter 2). Unfor-

Unfortunately, the CAM3 finite volume core has artificial damping, or fails to perfectly conserve angular momentum, and therefore does not create the superrotation observed in Titan’s deep atmosphere [Friedson et al., 2009]. The zonal acceleration of the winds is weighted by the cosine squared of the latitude, which centers it on the equator. The zonal forcing term is pressure dependent and decreases toward the surface. The forcing term becomes negative and the forcing switches from prograde to retrograde below the mean surface pressure of 146,594 Pa. When we add topography to the model, it affects the surface pressure and thus wind forcing. At lower pressures (higher elevations) the forcing term is prograde, thus we see prograde winds at high elevations in simulations with this forcing. At higher pressures (lower elevations) we get retrograde forcing, thus we see retrograde winds in basins in simulations with this forcing. Fig. 4.1 shows the ad hoc forcing at 170 m in our lowest model layer. Warm colors are prograde and cool colors are retrograde directions. Near Titan’s surface the magnitude of the forcing term is 10^4 smaller than the maximum tidal torques. These forcing terms are quite small compared to the oscillating tidal torques, however the tidal torque tends to cancel out over a Titan day, while this ad hoc forcing is constant in time.

4.3 Results

4.3.1 Dune elevation analysis

We have overlain the dune orientation directions from Lorenz and Radebaugh [2009] onto the topographic map provided by the Cassini Radar team [Lorenz et al., 2013] in Fig. 4.2. The topography in this map has been binned at 5 degree resolution. A histogram of the elevations of all Titan latitudes below 30 degrees, and those locations with dunes present shows that many dunes lie at higher elevations (Fig. 4.3). The mean elevation of the dunes compared with a 2575 km spheroid is -99 m. The mean elevation of equatorial region (+/- 30 degrees) is -183 m, and the mean elevation for all of Titan is at -388 m. It appears that Titan’s dune fields lie near high elevations, probably because the tropics themselves are at higher

elevation. It has also been proposed that wetlands, or moist conditions may suppress dune formation at low elevations [Neish and Lorenz, 2014]. The dunes likely lie in local basins in dry regions, as is the case with all sand seas on Earth spanning an area greater than 12,000 km² [Wilson, 1973] as well as the similarly large north polar sand seas on Mars [Tanaka et al., 2008]. Lorenz et al. [2013] argue that the dunes appear to be interrupted (deflected or truncated) by steep slopes, even if the change in elevation is small, but they are not affected by even large changes in elevation over long distances, i.e. low slopes. Our analysis disagrees with that of Le Gall et al. [2012] who argue that dune fields lie at low elevations. Reasons for the discrepancy between our analysis and that of Le Gall et al. [2012] could include different topographic maps or the resolution of our maps and comparisons. Their paper looks at each dune field in detail, while we investigate on a much larger scale. The goal of this paper is to investigate the dune forming winds with a GCM that runs at 4x5 degrees latitude x longitude resolution, so modeling the dune fields at higher resolution is beyond the scope of this study.

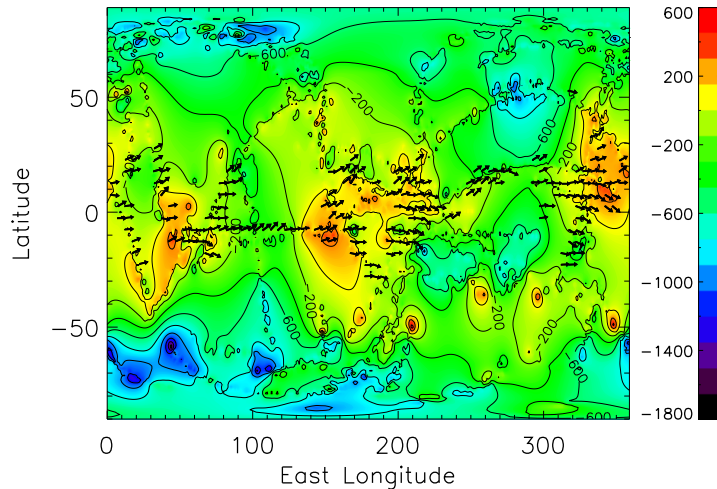


Figure 4.2: Topographic maps created by the Cassini Radar Team Lorenz et al. [2013] and reduced to 4x5 degree resolution overlain with dune orientations from Radebaugh et al. [2008].

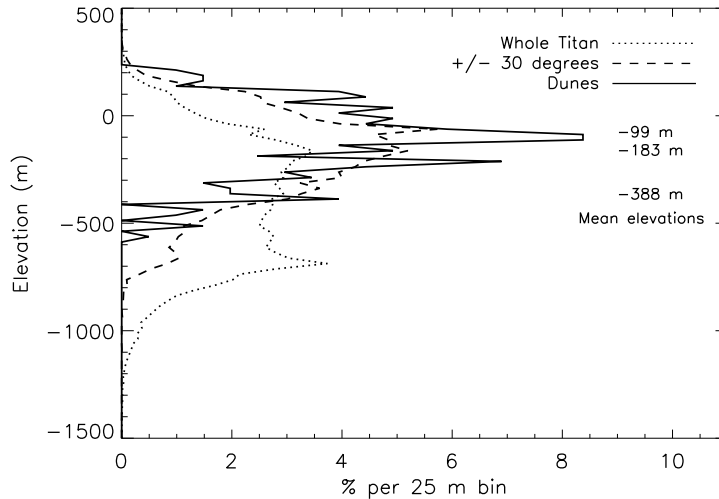


Figure 4.3: Hypsometry of grid cells at dune locations, less than 30 latitude, and of all Titan.

For a dune to form there needs to be sufficient sediment supply and suitable wind conditions. Along with topography, latitude clearly has a role in this as well. Fig. 4.4 shows the latitudinal and longitudinal distribution of the dune fields and topography. The solid lines in this figure are the number of dune fields. The dashed lines in Fig. 4.4 show the mean elevation in 1 degree bins. Surface zonal mean elevation is relatively high in the tropics and mid latitudes and low at the poles. The meridional mean surface elevation and number of dune fields seem to follow a similar pattern of highs and lows. Outside of the tropics, dunes are sparse. This could be due to the tidal forces from Saturn on Titan's winds, which tend toward the equator. Overtime, this could move sediment into the tropics. Alternatively, the mid and high latitudes may not have sufficient sand supply due to wet conditions.

4.3.2 GCM simulations of the near surface winds

We ran four simulations at 4x5 degree resolution to explore our two model additions mentioned above, topography and ad hoc forcing. The simulations are: 1) the base case model (bcm) , 2) the bcm with topography, 3) the bcm with ad hoc forcing, 4) the bcm

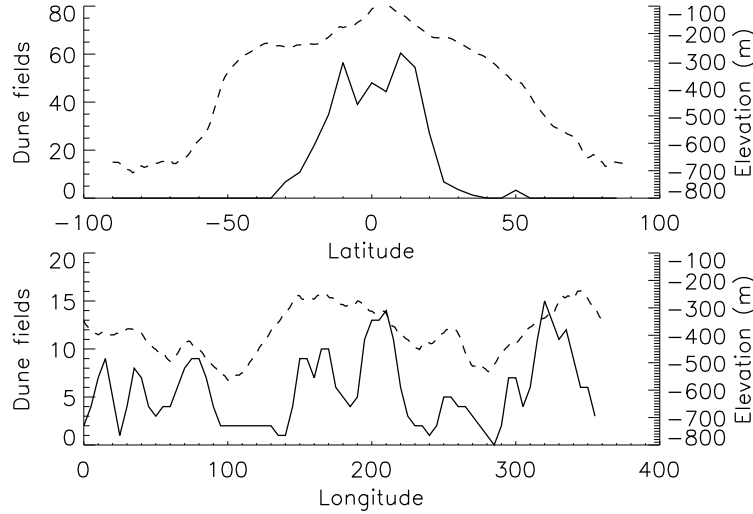


Figure 4.4: Profiles of the area weighted number of dune fields from Radebaugh et al. [2008] in 5 degree bins (solid) and the area weighted mean elevation in 1 degree bins (dashed) in latitude (top) and longitude (bottom).

with ad hoc forcing and topography. Fig. 4.5 shows the near surface zonal winds on Titan averaged over half an Earth year during the Northern hemisphere winter for each simulation. The near surface winds are the winds in our lowest model layer at 173 m altitude.

In our base case model we see few areas with prograde westerly near surface winds within 30° of the equator (Fig. 4.5, upper left). We do have bands of westerly winds in the midlatitudes that fluctuate seasonally, however they rarely lie within 30° of the equator. The tropics in this simulation have easterly winds which are inconsistent with the wind direction from the Huygens probe descent [Kazeminejad et al., 2007] as well as the dune forming wind theories discussed above. The Huygens probe reached a maximum horizontal velocity of 0.73 m/s toward the east in the last half kilometer of altitude. At 130 m above the surface, similar in height to our lowest model layer (173 m), the Huygens probe was moving at 0.3 m/s to the east. This is within the range of wind velocities in our base case simulation, but in the opposite direction. Analysis of the Huygens probe wind suggested that its velocity switched direction twice in the last 5 kilometers of descent, which is inconsistent with an

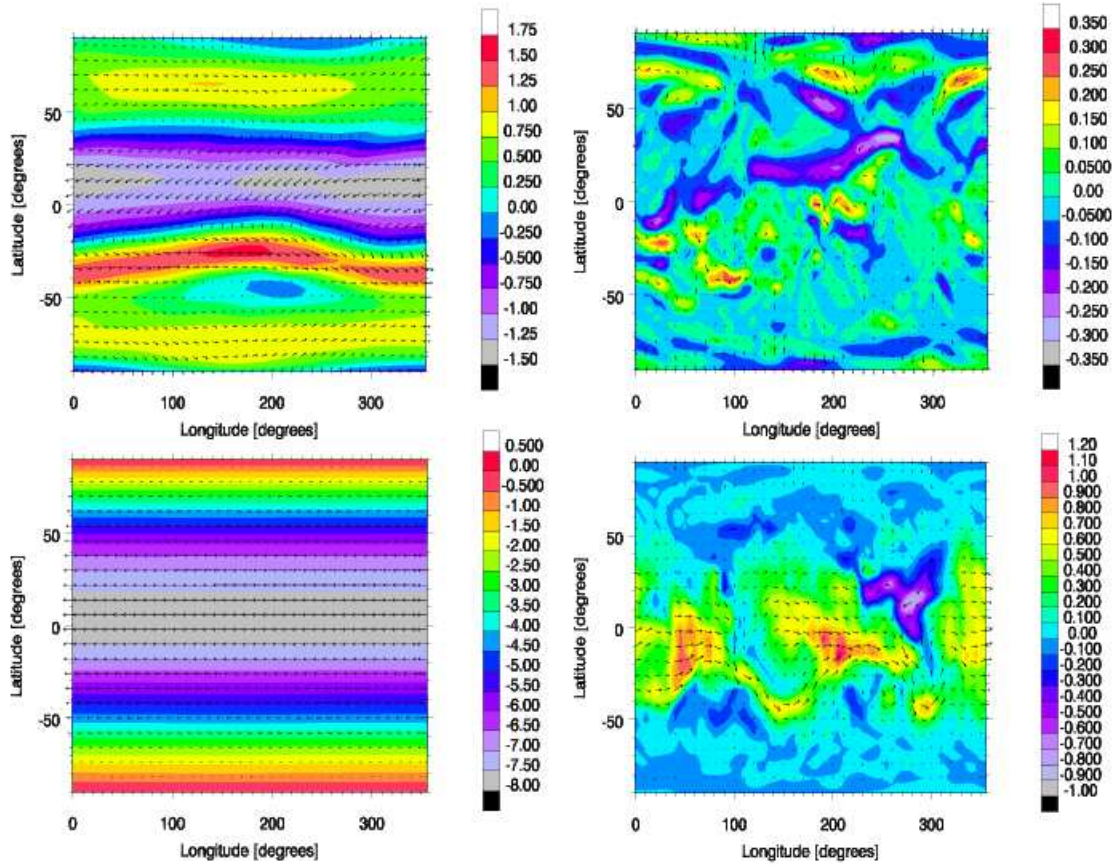


Figure 4.5: Zonal winds (colors) with wind vectors over plotted for the four simulations. The color bar indicates magnitudes for the zonal wind component. Note the change in scale. Clockwise from the upper left, they are: 1) the base case model (bcm), 2) bcm with topography, 3) bcm with topography and ad hoc forcing, 4) bcm with ad hoc forcing.

Ekman spiral, likely due to the weak Coriolis force [Tokano, 2007].

Adding realistic topography to our model drastically changes the near surface winds (Fig. 4.5, upper right). Topography reduces the magnitude of the winds by a factor of 5-20. However, the winds are also in a different direction in this simulation. Much of the tropics have prograde westerlies with average wind velocities up to 0.3 m/s. However, some areas of the tropics still have easterly winds with similar magnitudes. The change in direction to westerlies is more consistent with the Huygens probe direction. The Huygens probe velocities are near the high end of the velocities in this simulation. However, given the coarse grid used,

we cannot say that they are inconsistent.

The addition of the ad hoc torque (Fig. 4.5, bottom left) to our base case model creates easterly winds of several meters per second. These near surface winds are inconsistent with all of the dune forming wind theories, which require net westerly winds, in direction and magnitude as well as the Huygens descent probe tracking. The wind pattern in this simulation matches the equatorially peaked acceleration imposed on the model with the strongest winds at the equator falling off with increasing latitude.

The addition of topography and ad hoc acceleration has the interesting effect of forcing the winds retrograde in deep depressions and prograde at higher elevations. The forcing is pressure dependent and the surface wind in this simulation matches the pattern of ad hoc forcing in Fig. 4.1. The forcing is very weak, however it still dominates the surface wind. This simulation with both topography and ad hoc torque (Fig. 4.5, bottom right) shows large areas of westerly wind in the tropics with velocities up to 1.1 m/s. These winds are consistent in direction and velocity with the Huygens measurements. They also very closely match the direction of the dune orientations, as is discussed in section 3.3.

The wind patterns above are a snapshot from the northern winter on Titan. To get a sense of the wind pattern throughout the year, we have plotted downwind rose diagrams. Fig. 4.6 shows the downwind rose diagrams (wind directions from the center out) compiled over a Titan year at the equator and anti-Saturnian longitude (180°). This location was chosen for direct comparison with the gcm results of Tokano [2010]. The wind pattern for the base case model (upper left) is bimodal with the both directions being easterly and about 68° apart. This bimodality can be important for determining the type of dunes created. It has been shown that to create longitudinal dunes, nearly equal bimodal winds are required with divergence angles greater than 90 degrees [Rubin and Hunter, 1987]. However, Rubin and Hesp [2009] show that unimodal winds can create longitudinal dunes if the sediment is stabilized. There are currently no constraints on the stability of Titan's dune forming material. The simulation with topography (upper right) also has bimodal winds, but a large

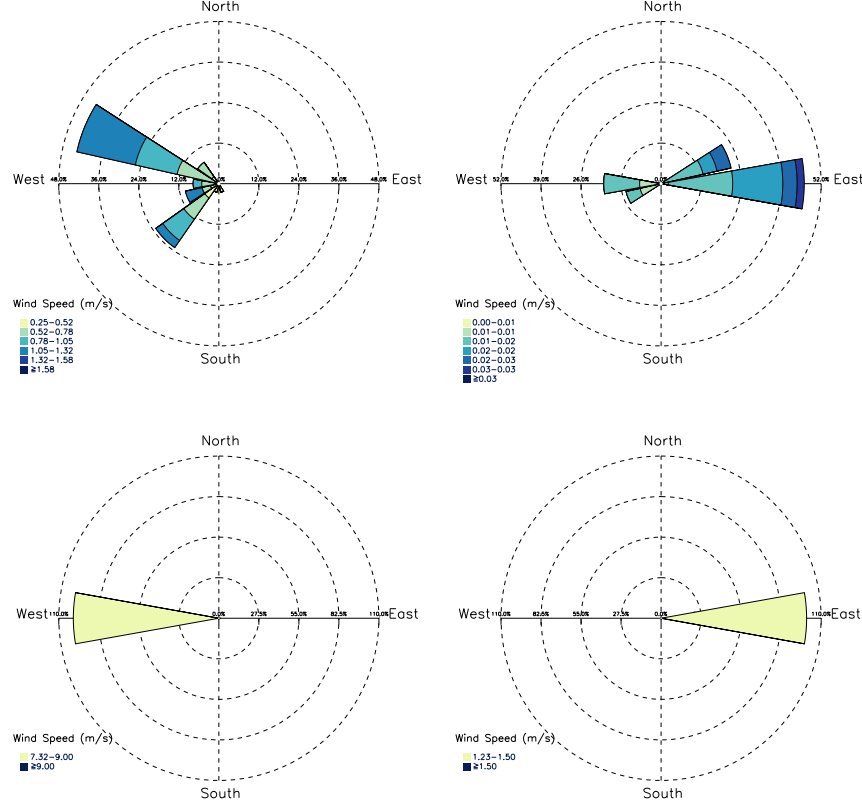


Figure 4.6: Downwind windrose diagrams (wind directions from the center out) for the anti-Saturnian longitude (180°) at the equator for one Titan year. Placement is the same as Fig. 4.5. Note the change in scale.

majority of the winds in this simulation are westerlies. The addition of the ad hoc forcing (Fig. 4.6, bottom panels) reduces the modality of the winds. The winds are constant in time due to the forcing.

However, it is clear from 4.5 that the wind direction and magnitude in the tropics is highly spatially variable. We also plot the same windrose diagrams for the Huygens landing site (10° S, 167° E) in Fig. 4.7 to demonstrate this variability. The simulations without artificial acceleration have the most spatial variation. The base case simulation with and without topography (top panels, Fig. 4.6 and 4.7) still show bimodal winds with important differences in wind speed and direction between the anti-saturnian point and Huygens landing

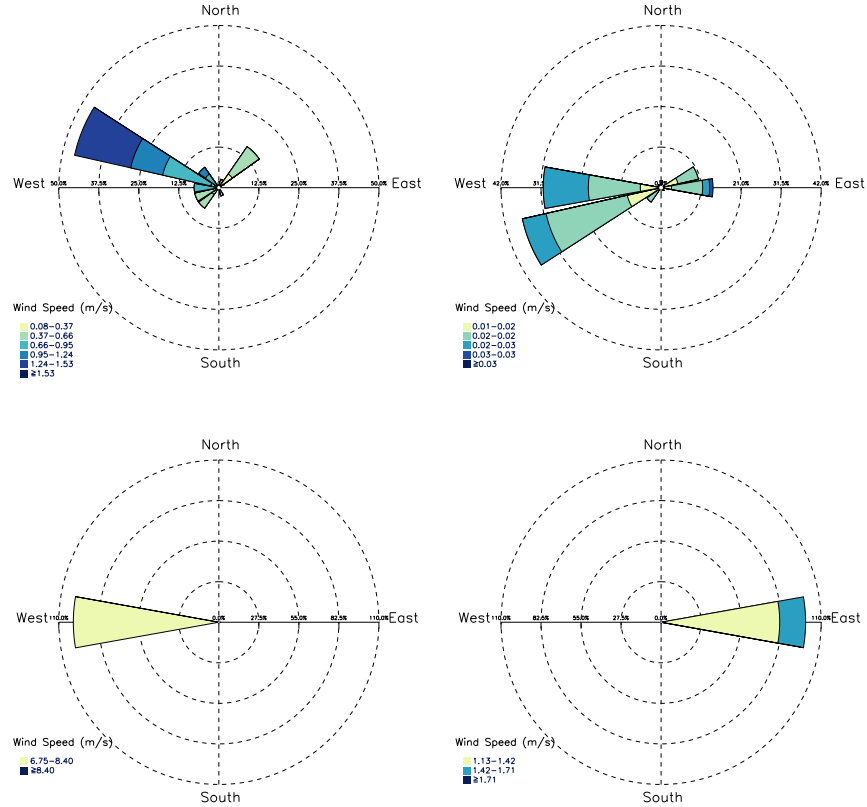


Figure 4.7: Downwind windrose diagrams (wind directions from the center out) at the Huygens landing site (10° S, 167° E) for one Titan year. Simulation placement is the same as Fig. 4.5. Note the change in scale.

site due to the wind interaction with topography. The forced simulations look very similar between these two locations.

Dune formation occurs due to a spatial or temporal decrease in the sand-carrying capacity of the wind in areas with strong enough wind that sand can be mobilized. Given Titan's fairly weak winds, a minimum condition needed for dune formation is sufficient wind speed to lift sand particles. Histograms of our simulated wind speeds at all locations (clear) and those with dunes (shaded) are shown in Fig. 4.8. We find that on average, dunes are present in grids with slightly higher wind speeds in our simulations, however the correlation is very low, 0.01-0.2. It is probable that the low resolution in our model is simply unable to

simulate the winds precisely enough to achieve a good correlation.

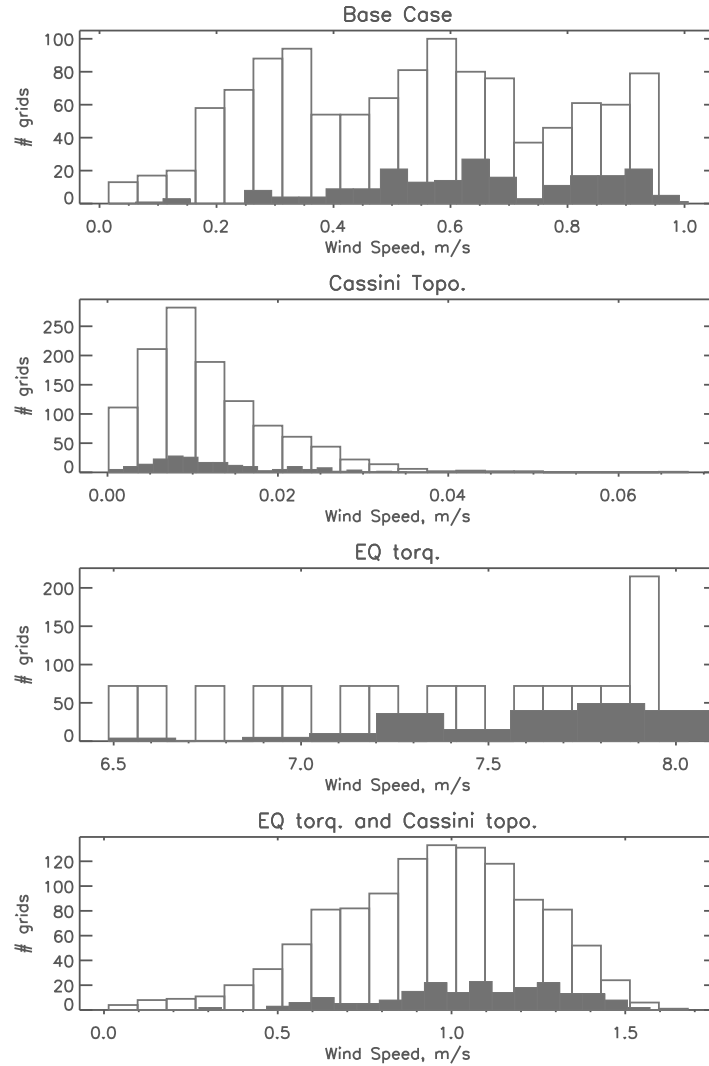


Figure 4.8: Average wind speed histograms of the simulations during one Titan year. Winds are averaged over an Earth year. Outlines indicate all of Titan, while the shaded region indicates grid cells with dunes.

4.3.3 Resultant drift direction and maximum gross bedform-normal transport

The resultant drift direction (RDD) is the mean direction of the potential sand transport over a period of time and is the direction that sediment tends to move (following the

terminology and method of Fryberger and Dean [1979]). The movement of sediment depends on the surface wind speed and the threshold speed for saltation. We calculate the RDD in a similar manner to Tokano [2008]. We assume that the friction speed scales logarithmically with altitude and is a function of our lowest model layer winds at 170 m above the surface. In Fig. 4.9 we plot the RDD for all dune locations in our four simulations as well as the mean dune crestline orientations at two different threshold speeds for saltation, 0 and 1 m/s. The speeds given are those of our lowest model mid layer, about 170 m above the surface.

The dune crestlines point in the direction of the expected average prevailing winds based on analysis of the dune interaction with topography. The base case and zonally forced simulations transport sediment in the opposite direction (towards the West) than the dune crestline orientations suggest. The simulation with topography improves the consistency between the crestline orientations and the simulated RDD. However, when we increase the threshold speed for saltation to 1 m/s, this simulation no longer transports sediment. The simulation with topography and ad hoc forcing transports sediment to the East, even at threshold speeds of 1 m/s, consistent with the dune orientations.

However, dune formation theory suggests that most dunes will orient themselves perpendicular to the direction of gross transport of sediment, which is typically different from the RDD [Rubin and Hunter, 1987, Fenton et al., 2014]. We can calculate the maximum gross bedform-normal transport (MGBNT) orientations for our simulated winds following the methods of Fenton et al. [2014]. Fig. 4.10 shows the MGBNT for our different model simulations in Titan's tropics. In general, our simulated winds would create dunes that are perpendicular to the linear dunes observed on Titan. However, the simulation with topography shows the most diversity in dune orientations, some of which are consistent with the dune crestline orientations on Titan.

We calculate the angle between the MGBNT and RDD. This angle is the "obliquity". For 0-15 degrees, the method of GBNT predicts the dunes will be longitudinal. For 15-75 degrees, the dunes will be oblique. For 75-90 degrees, the dunes will be transverse. At the

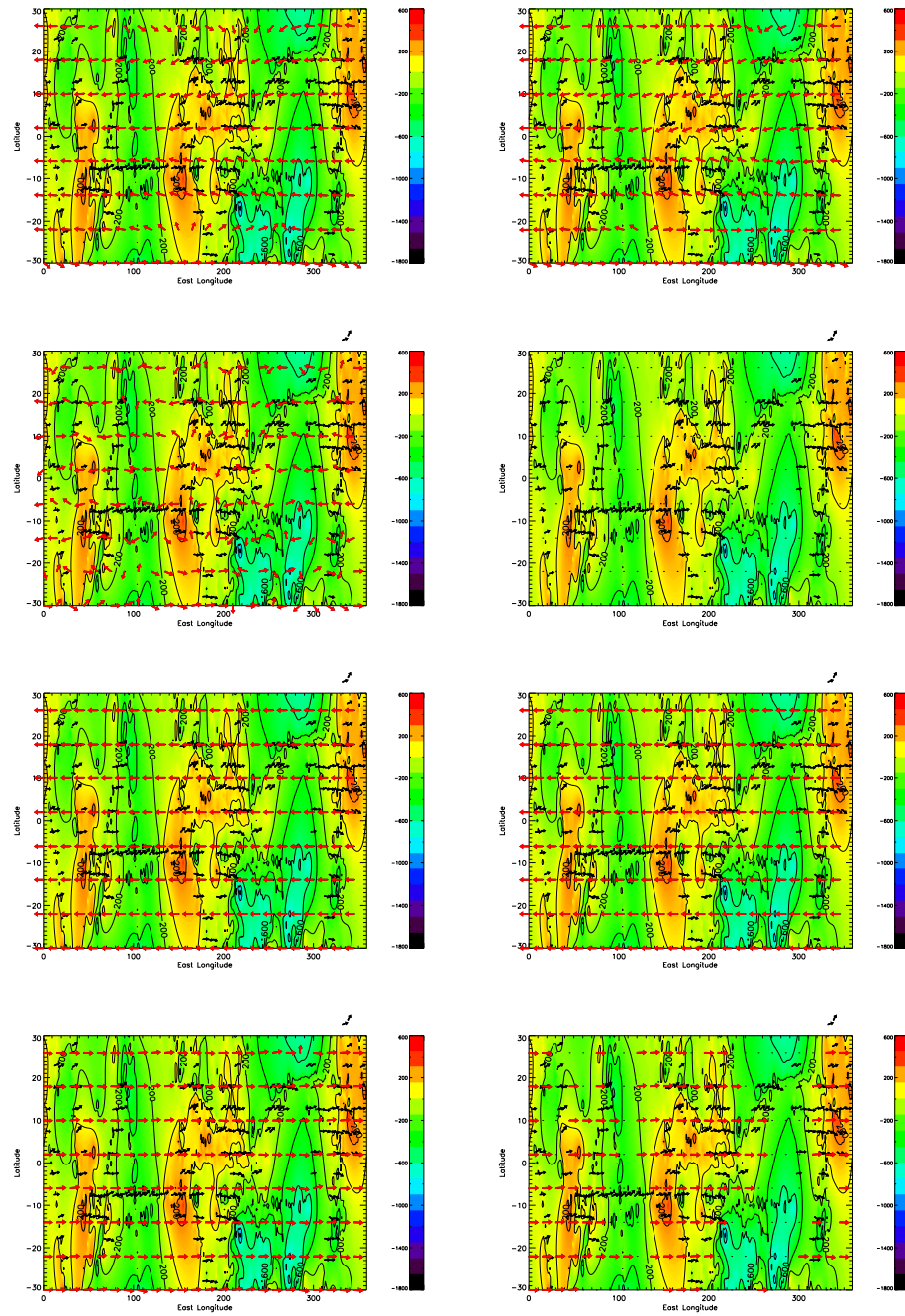


Figure 4.9: Plots of the dune crestline orientations (black arrows) as well as the RDD (red arrows) at 0 m/s (left) and 1 m/s (right) threshold speeds in Titan's tropics. The background colors are elevation contours. From top to bottom, the simulations are the base case model, bcm with topography, bcm with ad hoc forcing, and bcm with topography and forcing.

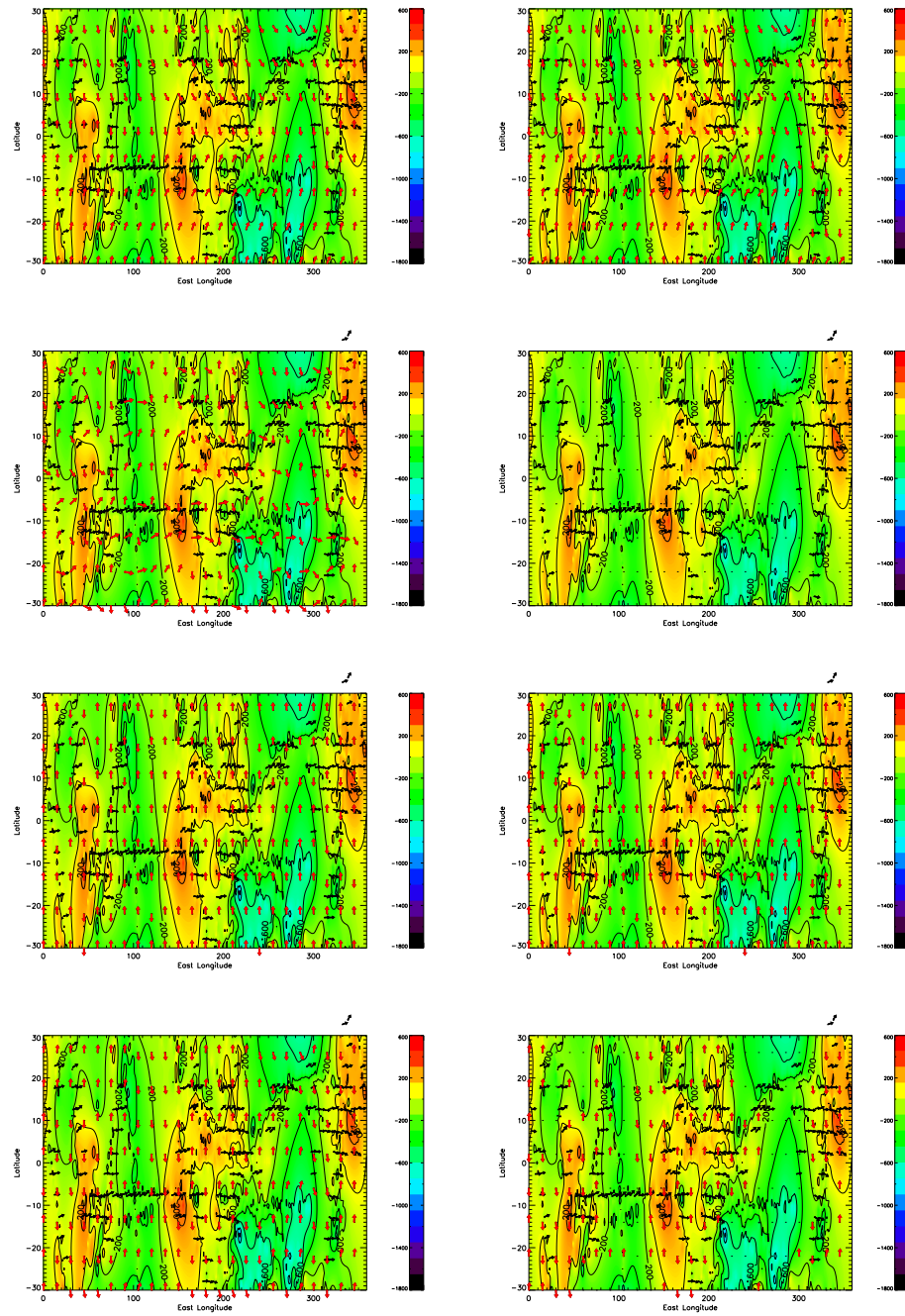


Figure 4.10: Plots of the dune crestline orientations (black arrows) as well as the MGBNT (red arrows) at 0 m/s (left) and 1 m/s (right) threshold speeds in Titan's tropics. The background colors are elevation contours. Simulation placement is the same as Fig. 4.9.

anti and sub-saturnian points, our simulations predict obliquity angles 75 and 90 degrees. Our simulation predicts transverse dunes on Titan. However, as was noted earlier, Rubin and Hesp [2009], found that unimodal winds in China created both transverse dunes where dune particles were free to move, and longitudinal dunes where the dune particles were stabilized due to a larger contribution of clay. This could suggest that Titan's dunes particles are somewhat stabilized.

4.3.4 Time analysis of surface winds

The prevailing theory of dune formation on Titan comes from Tokano [2010], who argues that strong westerly winds around equinox are responsible for creating the dunes. Throughout most of the year the wind speeds are less than 1.5 m/s and net easterly. However, around equinox, westerly zonal winds with speeds up to 1.8 m/s occur. By placing the saltation threshold speed for dune particles at 1.5 m/s, he is able to create dunes with westerly transported material. We investigate this directional time dependence in our model. In Fig. 4.11 we present the zonal winds throughout a Titan year at 5 different latitudes, -30, -15, 0, 15, and 30 degrees. Our simulations do not produce stronger zonal winds at or around the equator near equinox. The most seasonal diversity in the winds is seen in our base case simulation in which the winter 30 degree latitude has a zonal wind spike at the solstice.

4.4 Conclusions

We find that the inclusion of topography in our GCM changes the wind regime creating more westerly winds. The addition of topography and an artificial forcing term that depends on elevation creates a wind pattern in which the mean wind direction matches the dune orientations to a high degree. Furthermore, the forced simulation with topography is the only simulation with a mean net westerly resultant drift direction, consistent with the dune orientations. However, the addition of this ad hoc forcing reduces the modality of the

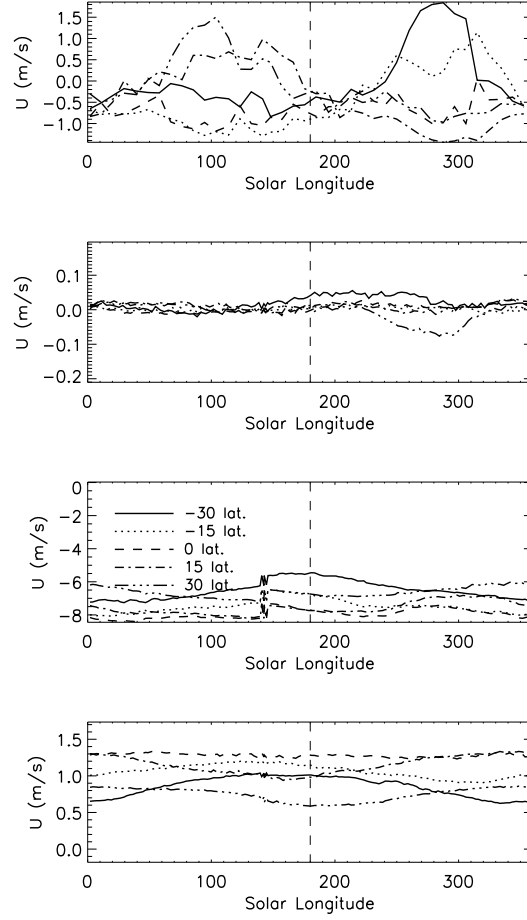


Figure 4.11: Zonal wind speed averaged over 3 earth months versus time for one Titan year in the four simulations. From the top down, they are the base case model (bcm), bcm with topography, bcm with ad hoc forcing, and bcm with topography and ad hoc forcing. Note the difference in scale.

wind, thus making it inconsistent with MGBNT theory. It is not certain how much of a constraint bimodality is for Titan's dune forming winds. Rubin and Hesp [2009] found that unimodal winds can also produce linear dunes when the sediment is locally stabilized. The stability of Titan's sediment is currently unconstrained. The angle between the RDD and MGBNT predicts the dune morphology. Our simulations predict transverse dunes, however the simulation with topography and no forcing is most consistent with longitudinal dunes at some locations.

This paper highlights the need for further work on Titan's dune forming winds. The dunes on Titan are responding to the local winds which are variable on a much smaller scale than we are currently able to model, especially if topography plays an important role in the dynamics. Thus much higher resolution GCMs or meso-scale models are required to resolve this interaction between the dune forming winds and topography.

Chapter 5

Conclusion

The objective of this dissertation is to characterize the physical properties and distribution of aerosols in Titan's atmosphere, and to determine their effects on the atmosphere. This is accomplished using a three-dimensional global circulation model with coupled aerosol microphysics. After characterizing the main haze and validating the model by comparing with observations, the GCM is applied to the formation of the detached haze layer and the dune forming surface winds on Titan, both being topics of interest in the field.

The second chapter establishes a best-guess set of microphysical properties that describe the aerosol in Titan's atmosphere based on sensitivity studies of the microphysical parameters. The important parameters, which control the size, shape, and abundance of the haze, are the production rate, fractal dimension, and electrical charge on the particles. The aerosol haze is comprised of aggregate particles with a fractal dimension of 2. The particles in the troposphere are larger with a higher fractal dimension likely due to condensation. The size of the aerosol particles is controlled by coagulation, which is inhibited if the particles are charged. A charge on the particles equal to 7.5 electrons/micron radius best fits the DISR observations of phase function and number density, and a production rate of 10^{-14} g/cm²/s best matches vertical extinction profiles in Titan's atmosphere. This production rate falls within the range of values published from both theoretical and laboratory investigations. However, it is three times lower than predicted by a one-dimensional model, which parameterizes vertical diffusion. This chapter also informs our understanding of Titan's north-south

albedo ratio and its seasonal cycle, despite a phase lag in the model. Changes in the aerosol size and number density create a wavelength-dependent albedo variation between Titan's hemispheres.

Despite much effort, the model was unable to reproduce the superrotating winds observed on Titan. This is likely because the CAM3 finite volume core does not conserve angular momentum. To get around this issue and still have realistic dynamics in the model, I employ ad hoc forcing to the zonal winds. The forced dynamics permit large meridional temperature gradients consistent with observations to form in our simulations.

Chapter 3 presents a formation mechanism for Titan's detached haze layer based on a balance between the vertical winds and particle fall velocities. The summer hemisphere on Titan has upwelling winds associated with a pole-to-pole Hadley cell that transport aerosol particles upward. However, high in the atmosphere, aerosol movement is net downward, dominated by their fall velocities. At some point, these two velocities are equal and the particles are suspended, forming the detached haze layer. An analytical model is used to demonstrate this formation mechanism and match the vertical extinction profiles from West et al. [2011]. The change in altitude of the detached haze layer around equinox is due to a change in the vertical winds associated with the overturning Hadley cell. Titan's detached haze layer imaged by Cassini between 2005 and 2012 merged with the main haze at the end of 2012, consistent with the predictions from our model. These simulations suggest the detached layer will reform at its original altitude around 550 km, between mid 2014 and early 2015.

Finally, chapter 4 analyzes how the addition of topography and ad hoc acceleration in our model affects the surface winds. Topography changes the direction and magnitude of surface winds, making them more aligned with dune crestline orientations on Titan. The addition of ad hoc acceleration by itself did not produce Titan-like surface winds. However, coupling the forced dynamics with the topography produced winds consistent with the dune orientations on Titan.

5.1 Future Work

This thesis opens up several areas for future work. Important additions to this research would include: 1) parameterization of clouds and precipitation in the model, 2) adding condensation to the aerosols to investigate the changing wavelength dependence of the optical depth in the troposphere, 3) analyzing Titan's detached haze observations at many latitudes to get a more complete spatial and temporal picture of the detached haze, and 4) running the model with reversed periapsis, i.e. Titan is closer to the sun during the northern summer, in order to test wind direction and methane transport dependence on orbital properties.

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Appendix A

Appendix

A.1 Bin sizes

We used 30 aerosol bins with the fractal properties shown below in Table A.1. N_{mon} is the number of monomers. For aerosols with a total radius above 50 nm, the monomer size is 50 nm. DF is the fractal dimension, R_s is the equivalent mass spherical radius, R_f is the fractal radius. Our model calculates three different radii for each mass bin. The first is the equivalent mass sphere radius, r . This is the radius the fractal aggregate particle would have if it were condensed down to a sphere of the same mass. For any equations in the model which use particle radius to calculate a mass, we use the equivalent mass sphere radius. We also calculate the fractal radius, r_f , which depends on the degree of packing of the aggregate. This radius is used in equations in which the particle size is important, such as coagulation. The last radius we calculate is the mobility radius, r_m . The mobility radius, discussed below, takes into account the porosity of the fractal aggregates when falling and is slightly smaller than the fractal radius. This radius is used in fall velocity equations. See Vainshtein et al. [2004] for more details on the the mobility radius.

A.2 Coagulation equations

Coagulation happens through two processes, Brownian motion, and gravitational sedimentation with the coagulation kernel equal to the sum of these two coefficients. Gravita-

Table A.1: The particle size bins used in the model and their fractal properties.

Bin	N_{mon}	DF	$R_s (\mu m)$	$R_f (\mu m)$	$R_m (\mu m)$
1	1.0E+0	3.00	0.002	0.002	0.002
2	1.0E+0	3.00	0.003	0.003	0.003
3	1.0E+0	3.00	0.004	0.004	0.004
4	1.0E+0	3.00	0.005	0.005	0.005
5	1.0E+0	3.00	0.007	0.007	0.007
6	1.0E+0	3.00	0.009	0.009	0.009
7	1.0E+0	3.00	0.013	0.013	0.013
8	1.0E+0	3.00	0.017	0.017	0.017
9	1.0E+0	3.00	0.023	0.023	0.023
10	1.0E+0	3.00	0.031	0.031	0.031
11	1.0E+0	3.00	0.042	0.042	0.042
12	1.5E+0	2.00	0.058	0.062	0.059
13	3.8E+0	2.00	0.078	0.098	0.089
14	9.5E+0	2.00	0.106	0.154	0.136
15	2.4E+1	2.00	0.144	0.244	0.214
16	6.0E+1	2.00	0.195	0.385	0.340
17	1.5E+2	2.01	0.265	0.606	0.543
18	3.7E+2	2.01	0.360	0.951	0.866
19	9.3E+2	2.02	0.488	1.478	1.368
20	2.3E+3	2.04	0.663	2.259	2.138
21	5.8E+3	2.06	0.899	3.344	3.189
22	1.5E+4	2.11	1.221	4.674	4.514
23	3.6E+4	2.20	1.657	5.954	5.814
24	9.1E+4	2.33	2.249	6.680	6.585
25	2.3E+5	2.53	3.052	6.610	6.564
26	5.7E+5	2.75	4.142	6.227	6.212
27	1.4E+6	2.92	5.621	6.396	6.394
28	3.5E+6	2.99	7.629	7.752	7.752
29	8.9E+6	3.00	10.35	10.36	10.36
30	2.2E+7	3.00	14.08	14.08	14.08

tional coagulation, or coalescence, is important for large particles and depends on the fall velocities and particle radii as well as the collision efficiency. It is not likely to be important on Titan since the particle sizes are relatively small. The equations that describe the coagulation in the model are as follows. The thermal (Brownian) coagulation coefficient is calculated by solving:

$$K_{br} = \frac{4\pi R_p D_p}{\left[\frac{R_p}{(R_p + del)} + \frac{4D_p}{\sqrt{(v_{th1}^2 + v_{th2}^2)} R_p C_{stick}} \right]} \quad (A.1)$$

where R_p is the sum of the two particle fractal radii, D_p is the sum of the diffusion coefficients, del is a function of the aerosol mean free path, v_{th} is the thermal velocity of the particles and a function of the temperature, and C_{stick} is the sticking coefficient. Where the size of the particle is important in determining if other particles collide with it, the fractal

radius is used. This includes the equations for coagulation and the Knudsen number.

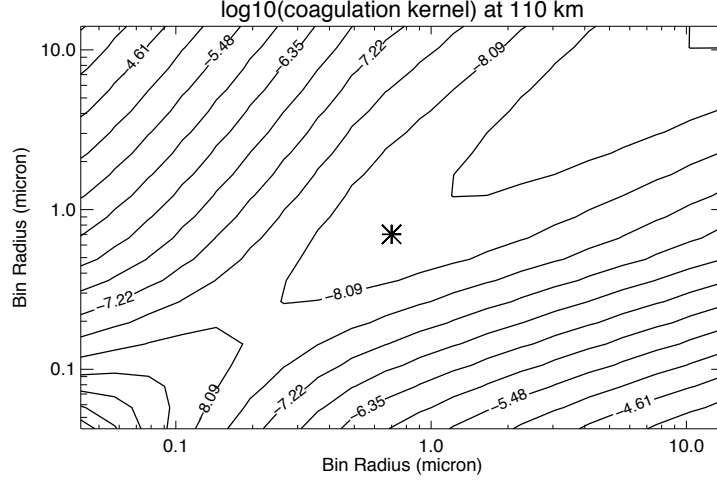


Figure A.1: The coagulation kernel of our best fit simulation with a charge to radius ratio of 7.5 electrons per micron. The asterisk symbol indicates the effective radius of our particles at 100 km.

This coagulation coefficient falls into two regimes based on the mean free path of the aerosols in the atmosphere. Where the mean free path is large, λ is large, and we are in the kinetic limit. Here the coagulation coefficient simplifies to $K_{br} = \pi R_p^2 (v_1^2 + v_2^2)^{1/2} C_{stick}$. On Titan, this occurs above 80 km where the mean free path of the air molecules exceeds a micron [Toon et al., 1980]. At about 50 km the mean free path becomes $0.1 \mu\text{m}$ and we enter the diffusion, or Stokes, limit. Here the coagulation coefficient simplifies to $K_{br} = 4\pi R_p D_p$.

In Titan's atmosphere, it is thought that significant charging due to cosmic rays and precipitating electrons inhibits the coagulation of aerosol particles [Borucki et al., 1987, Toon et al., 1992]. This inhibition is manifested in the sticking coefficient, C_{stick} , which is calculated as:

$$C_{stick} = e^{\left[\frac{-r_{f1} r_{f2} p_r^2 e^2}{kT(r_{f1} + r_{f2})} \right]} \quad (\text{A.2})$$

Here r_1 and r_2 are the fractal radii of the aerosols, p_r is the charge to radius ratio, e

is the electron charge, k is the Boltzmann constant, and T is temperature. We set p_r to 0 - 15 (assuming r is in micrometers). Toon et al. [1992] found that for particles with a few electrons per $0.1 \mu\text{m}$, coagulation is inhibited for particles around $0.1 \mu\text{m}$. Our best guess case has a charge to radius ratio of 7.5 electrons per micron however, allowing our particles to grow much larger than $0.1 \mu\text{m}$ [Barth and Toon, 2004]. Figure A.2 shows our coagulation kernel for different sized particles in our best guess simulation, as well as the particle size at 100 km.

A.3 Fall velocities

To include the fractal effect on the fall velocity of the aerosols CARMA follows the method of Vainshtein et al. [2004]. We define a mobility radius, r_m , which takes into account the porosity of the fractal aggregates.

$$r_m = \Omega r_f \quad (\text{A.3})$$

Here omega is the ratio of the drag of a permeable particle over the drag of an impermeable particle and it is always less than one, see Vainshtein et al. [2004] for details. Using this radius tends to increase the aerosol fall velocities compared to using the fractal radius. They are not increased by a large amount compared to the fall velocity for fractals however, and never approach the velocity of spheres. Equations used in the fall velocity of the aerosols use this mobility radius. These include the Reynolds number, diffusion equation, and Stokes fall velocity.

CARMA calculates the fall velocity, V_g , as the following. In the Stoke's limit, in which the mean free path is small compared to the particle size.

$$V_g = \frac{2(r_m^2 \rho g C)}{9\mu} \quad (\text{A.4})$$

When the mean free path is large compared to the particle size we are in the kinetic limit and the equation takes the following form:

$$V_g = \beta \frac{\rho g r_m}{2n_a} \sqrt{\frac{\pi}{2m_a k T}} \quad (\text{A.5})$$

In equation /refvg2, μ is the dynamic viscosity of air, r_m is the particle mobility radius, ρ is the particle density, n_a is the number density of air molecules, and m_a is the mass of an air molecule. C is the slip correction factor, which is extrapolated from the mean free path and is calculated as

$$C = 1 + \frac{\lambda}{r_m} [1.257 + 0.4e^{(\frac{-1.1r_m}{\lambda})}] \quad (\text{A.6})$$

where lambda is the mean free path of air. When the particles are larger than the mean free path, the fall velocities depend on particle size, viscosity and density. However, when the particles are smaller than the mean free path, the velocity only depends on the particle size and air density.

A.4 Optical depth

We use code provided by Rannou Pascal and described in Botet et al. [1997] to calculate the optical properties of the aerosols. This code uses Mie theory for spherical particles and a mean field approximation for aggregates of spherical monomers. From these optical properties and the aerosol abundances, we can calculate optical depths from the aerosols in our model and the aerosol heating rates.

$$\tau_a(\lambda) = \int_0^{\infty} \beta_e(\lambda, z) dz \quad (\text{A.7})$$

In equation A.7, τ_a is the aerosol optical depth. β_e is the extinction coefficient which depends on wavelength and altitude. These calculated optical depths can be compared to the observations to validate the model as in Figure 2.7. Fractal aggregates have much larger extinction coefficients at short wavelengths compared with spheres. The fractals and spheres are more comparable in their longwave absorption coefficients however. Figure A.2 displays

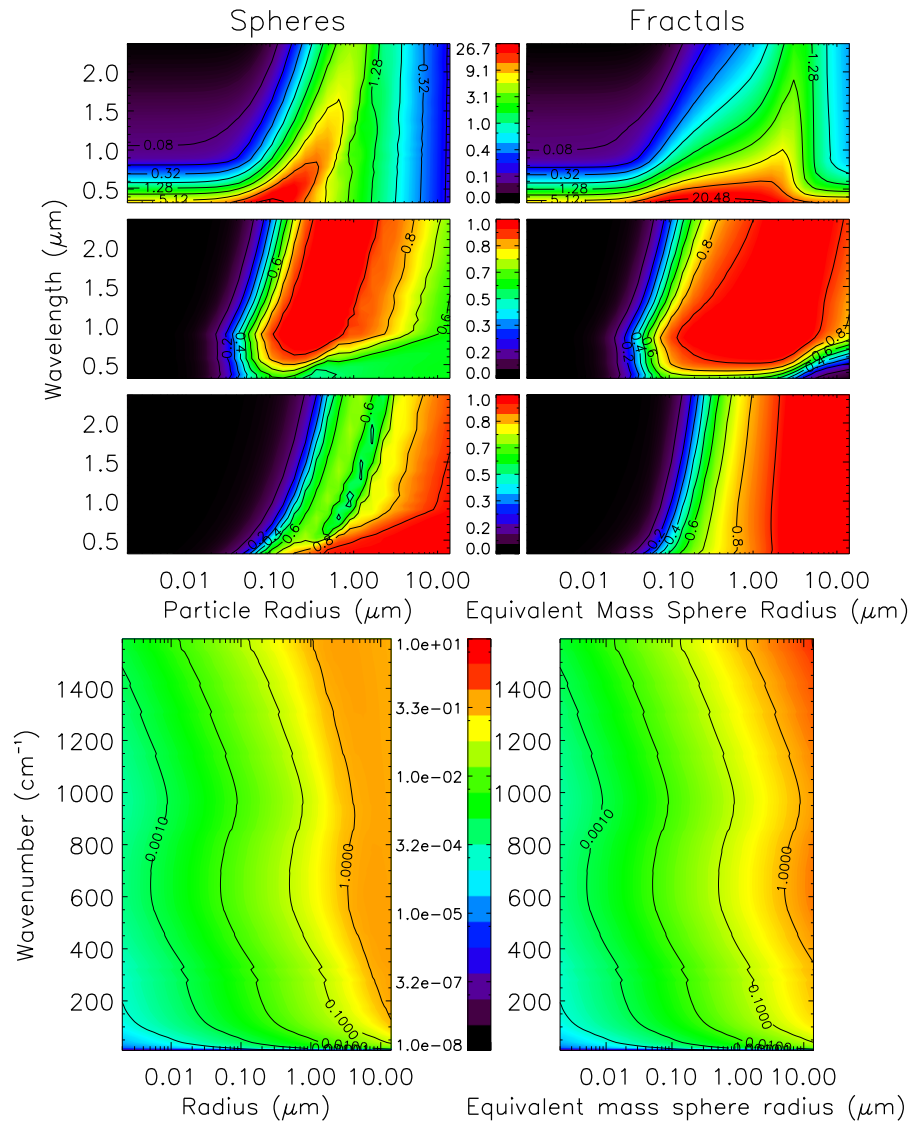


Figure A.2: Top, the Q extinction efficiency (1st row), single scattering albedo (2nd row), and asymmetry parameter (3rd row) for spherical (left) and fractal (right) particles used in the shortwave radiation code (300 nm to 2.5 microns). Bottom, the absorption coefficients of spherical and fractal aerosols used in the long-wave radiation code in CAM (wavenumbers 10 - 1500).

the optical properties of the spherical and fractal aerosols described above over a broad range of wavelengths from the near UV to IR.

A.5 Radiation code investigation of fractal dimension and monomer number

The fractal aggregate structure of Titan aerosols has been shown necessary to match the geometric albedo, wavelength dependence of aerosol optical depth, and phase function in the atmosphere of Titan [Cabane et al., 1993, McKay et al., 2001, Rannou et al., 1995, Tomasko et al., 2008b, Lavvas et al., 2010]. The fractal structure of aerosols is described by three main properties: 1) the monomer size, which is the size of the smallest particle from which the fractals are made, 2) the number of monomers, and 3) the fractal dimension. The fractal dimension is a measure of the compactness of the fractal aggregate. A fractal dimension of 1 would be like beads on a string, where the beads are the monomers. A fractal dimension of 3 is a sphere of overlapping monomers with the same mass and radius as a solid sphere. A fractal dimension of 2 is somewhere in between, similar to a snowflake.

Titan models have different ways of treating the fractal dimension in their analysis. Tomasko et al. [2008b] fixes the fractal dimension at 2 and adjusts the monomer size and number of monomers to match their data. Lavvas et al. [2010] have a production zone with a fractal dimension of three and below that the fractal dimension is 2. They closely fit the slope of the aerosol optical depth with wavelength above 80km, but have a poor fit in the lower atmosphere. Wolf and Toon [2010] have a unique method of making the fractal dimension a function of aerosol size with the rational that a few monomers sticking together would be more like a beads on a string than a snowflake, thus having a fractal dimension of less than two. In their model, larger aerosols, however, will be more compact and have a fractal dimension closer to 3. Cabane et al. [1993] indicate that the fractal dimension depends on the conditions in which the aerosol forms and finds a range of fractal dimensions between 1.75 and 3 depending on these conditions. Cabane et al. [1993] also identify two main regions, that of monomer growth where the fractal dimension is 3 and below that where the aerosols coagulate through ballistic collisions with a fractal dimension of 2. Generally, most models use a value around 2 for fractal dimension which does a good job of reproducing

the geometric albedo on Titan [Rannou et al., 1995, McKay et al., 2001].

We have used the mean field approximation code [Botet et al., 1997] as a stand alone radiation code to investigate the aerosol shape change in the lower atmosphere with the caveat that our results are not taking into account any change in the indices of refraction. This is the same code we use to calculate the aerosol optical properties in the GCM radiative transfer code. In Figure A.5, we explore the effect of fractal dimension and number of monomers on the slope of the wavelength dependence of the aerosol optical depth. We use a chi squared analysis of the parameter space to identify the best fit to the aerosols. Both Tomasko et al. [2008b] and Lavvas et al. [2010] use 3000 monomers as their best fit. We explored 2^0 to 2^{18} monomers by factors of 4, with our best fit having 4096 monomers. We explored fractal dimensions from 1.6 to 3.0 with fractal dimensions of 2.0, 2.4, and 2.8 being the best fits to altitudes above 80 km, 30-80km, and below 30 km respectively (Fig. A.5). Using the larger fractal dimensions (2.4 and 2.8) gives much better fits to the observations below 80 km than does the model with fractal dimensions of 2.0.

There are two main problems with the approach used in Fig. A.5. First, we modeled the aerosol optical properties of one size particle. The particles are not a single size, there is a size distribution. However, at any given altitude, one size should dominate as discussed in section 3.4. Second, as the fractal dimension approaches three the mean field approximation used in this model treats the particle aggregate as a collection of spheres, not a single sphere. As the monomers get compacted the monomer size could grow. If there is significant condensation from volatiles, they could form spherical droplets that should be treated as a single monomer possibly with different optical properties for the core and shell. Despite this uncertainty, modeling the aerosols with the mean field approximation is informative. The change slope of the wavelength dependence of the optical depth can be reproduced by only changing the fractal dimension of the aerosols.

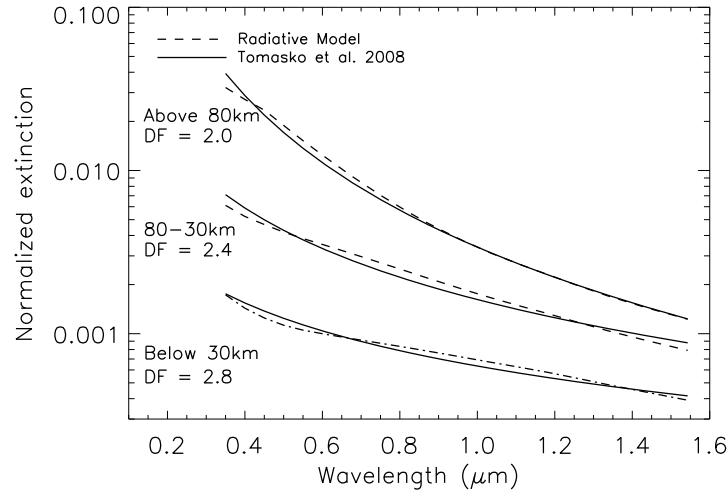


Figure A.3: Best fit fractal dimensions which match the slope of the wavelength dependence of the optical depth at each layer of the lower atmosphere identified by Tomasko et al. [2008b].

A.6 Image details

All images used in this paper were taken by the Cassini Imaging Science Subsystem. The left panel in Fig. 1.1 is image N1715034797 (CB3, 938 nm) taken on May 7th, 2012. The right panel in Fig. 1.1 is image W1715032186 (red, 650 nm). The left panel in Fig. 1.2 is a true color composite image produced from images N1712583549 (blue, 445 nm), N1712583582 (green, 562 nm), and N1712583615 (red, 650 nm). The images used in this composite were taken on April 9th, 2012. The right panel in Fig. 1.2 is image N1629620703 (blue, 44nm) taken on August 22nd, 2009. Fig. 3.1 is image W1681908348 (blue, 445 nm) taken on April 9th, 2012.