Observations of the global haze redistribution on Titan from 2006 to 2015 with OSIRIS at Keck

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Abstract

We observed Titan with the OH Suppressing InfraRed Imaging Spectrograph (OSIRIS) at the W. M. Keck observatory from 2006 through 2015 using adaptive optics. The sunlight scattered by atmospheric haze was spatially resolved in the $2.0\ \mu\text{m}$ (K) band window, and the spectra were analyzed with a radiative transfer model to determine the vertical (altitude) and meridional (latitudinal) variation in the haze distribution over this time period. This study complements recent work by Karkoschka (2016) in the season of observations, in the time span and sampling interval, in wavelength coverage and spectral resolution, as well as in the radiative transfer methodology and analysis. We observe the largest meridional gradient in haze opacity above 20 km toward the northern hemisphere in January 2010. Individual observations can show significant deviations from a relatively smooth linear gradient in haze across the entire disk. The variation in haze below 20 km is rarely well-described by a simple model and there is a systematically smaller amount of haze opacity retrieved from the Equator to 10°S when observing the disk with a sub-observer longitude near 150°W . This correlation with longitude suggests one of the following; a localized decrease in haze scattering, a localized increase in gas opacity, or a systematic over-estimate of the surface albedo in this region.

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1. Introduction

The hydrocarbon hazes on Titan are one tracer of the overturning circulation (e.g., Smith et al., 1981; Lorenz et al., 2001; Rannou et al., 2004; Larson et al., 2015). Together with the location and frequency of cloud formation, the distribution and evolution of hazes provide insight into the photochemistry and dynamics of the atmosphere (e.g., Strobel, 1974; Yung et al., 1984; Wilson and Atreya, 2004; Lavvas et al., 2008b; Krasnopolsky, 2009; Willacy et al., 2016). Podolak et al. (1979) noted the general acceptance of so-called “Axel dust” in the upper atmosphere of Titan as they followed the early laboratory experiments of Khare and Sagan (1973) in trying to understand the optical and chemical properties of this material. Allen et al. (1980) proposed a photochemical theory of the production of the polymeric compounds. Early models of the solar phase angle variations of Titan’s geometric albedo (Rages and Pollack, 1980) were used to determine the optical properties and vertical distribution of this material on a global scale. Voyager observations would provide the first resolved look at the hazes on Titan (Smith et al., 1981, 1982).

The properties of the sunlight that is scattered by hazes in the atmosphere is dependent on the shape and size distribution of particulates (Rages et al., 1983; McKay et al., 1989; Tomasko et al., 2008; Doose et al., 2016; Mishchenko et al., 2016), which vary spatially with geographic location as well as vertically, in altitude above the surface (e.g., Rages and Pollack, 1983; Lorenz et al., 2001; Anderson et al., 2008). Toon et al. (1992) presented a model to explain the hemispheric asymmetry that was observed by Voyager as a result of the Hadley-type motions above 300 km. These models were used to simulate the time-dependent production of haze, in order to reconcile the predicted geometric albedo of Titan with observations (Hutzell et al., 1993), before they were extended to two dimensions in order to interpret brightness variations observed with both season and latitude (Hutzell et al., 1996). The state of knowledge about aerosols on Titan, prior to the arrival of Cassini, was reviewed by McKay et al. (2001). Recent microphysical models of the vertical distribution of haze (Lavvas et al., 2008a) build on treat the photochemical processes that turn atmospheric methane gas into the larger hydrocarbon molecules as well as the details of the condensation process through heterogeneous nucleation and subsequent coagulation. The haze distributions in these models are solved self-consistently with the atmospheric thermal structure and are in agreement with the properties measured by the Descent

Imager/Spectral Radiometer (DISR) on the Huygens probe (Tomasko et al., 2005, 2008; Lavvas et al., 2011). While the measurements made by instruments on Cassini and Huygens provide the most detailed constraints for our understanding of the vertical structure of haze, the hazes on Titan are neither uniformly distributed nor static. The role of the seasonally changing insolation, and the resulting impact on the photochemistry that is related to haze formation, is not well constrained against the role of transport by large scale dynamics (e.g., Rannou et al., 2002), suggesting that additional information about the temporal evolution of hazes is valuable for evaluating the photochemistry and dynamics on Titan.

The *Hubble Space Telescope* (*HST*) was used to measure latitudinal variation in the size of aerosol (Karkoschka and Lorenz, 1997), and subsequent observations with *HST* revealed a stronger hemispheric contrast than expected (Lorenz et al., 1999) with changes in the brightness distribution over seasonal timescales and large scale albedo variations near the pole (Lorenz et al., 2001, 2004). A 3-dimensional map the haze in the lower atmosphere (in 1996), from ~ 100 km to the surface was assembled using 6 *HST* filters in the 888 to 953 nm range (Young et al., 2002), showing an upper atmospheric aerosol (above 16 km) that peaked near the equator and fell off toward both poles, and a lower atmospheric haze (below 16 km) that shows peak optical depths at both northern and southern mid-latitudes (near $30\text{--}40^\circ$).

These haze asymmetries can also be observed from telescopes on Earth in the near-IR (e.g., Gibbard et al., 1999; Roe et al., 2002; Ádámkovics et al., 2004, 2006). The Visual and Infrared Mapping Spectrometer (VIMS) on Cassini (Brown et al., 2004) has been used to measure the distribution of haze from 2004 through 2008, finding both gradual gradients as well as stepwise discontinuities in the haze (e.g., Rannou et al., 2010; Penteado et al., 2010). Spatially discrete features such as polar haze enhancements have also been observed with VIMS (Rannou et al., 2012), while a tropical haze band was observed using the Imaging Science Subsystem (ISS), together with VIMS and the Cassini Composite InfraRed Spectrometer (CIRS) (de Kok et al., 2010).

The high-altitude detached haze layers were initially observed by the Cassini ISS to be significantly higher (above 500 km) than measured by Voyager (below 400 km; Rages and Pollack, 1983). Near equinox, dramatic changes in vertical structure of the hazes were observed, with the altitude of the upper detached haze layer dropping from over 500 km down to below 380 km between 2007 and 2010, thereby connecting the Cassini observations

with Voyager measurements from the previous seasonal cycle (West et al., 2011). The evolution of the detached photochemical haze layer is one observational constraint in evaluating our understanding the relative importance of the haze microphysics, as regulated by variations in the photochemistry that are driven by insolation, compared to the role of global dynamics, wave structures, and winds. Meier et al. (2000) used *Hubble Space Telescope (HST)* observations in 1997 and 1998 to tentatively identify banded (zonal) haze structures close to the surface. Starting from the same epoch, Gibbard et al. (2004) measured the $2\mu\text{m}$ haze opacity using speckle and adaptive optics imaging at the W. M. Keck Observatory, finding the Southern haze opacity to be decreasing by a factor of about 2 from 1996 through 2004, with equatorial hazes staying relatively constant. Recently, Karkoschka (2016) presented an analysis of *HST* Space Telescope Imaging Spectrograph (STIS) observations of Titan from five years between 1997 and 2004. Those observations at 543 and 990 nm wavelengths were used to show that there were two separate seasonally varying components of atmospheric haze opacity above and below $\sim 100\text{ km}$ altitude. The extrema for the observed variation in haze opacities occurred in the earliest measurements in 1997, and there was a 2 year time period from 2002 through 2004 where the hemispheric asymmetries reversed in low, mid, and high latitudes.

Constraining the spatial variation in hazes closest to the surface, using wavelength windows that are most sensitive to the surface, is important for determining atmospheric properties in the boundary layer. Spatial variation in boundary layer methane is related to the source region(s) of methane (Ádámkovics et al., 2016), however, degeneracies between methane absorption and haze scattering in retrieval algorithms complicate an unambiguous determination of the methane abundance. Placing constraints on the low altitude haze opacity can therefore inform the retrievals of the methane abundance in the boundary layer. Here we present observations of haze variation that are complementary to a recent study by Karkoschka (2016). Our 14 observations, from early 2006 through mid 2015, are presented in Section 2. As well as covering a different season on Titan, our observations are in a longer wavelength band at $2\mu\text{m}$, as well as at higher spectral resolution. These near-infrared, ground-based observations of the haze evolution on Titan are interpreted with a radiative transfer model, discussed in Section 3. We present our results of fitting the haze variations in Section 4 and discuss the results in Section 5.

113 2. Observations

114 Titan was observed from January 2006 through July 2015 with the OH
 115 Suppressing InfraRed Imaging Spectrograph (OSIRIS) at the W. M. Keck
 116 Observatory. OSIRIS is a near-IR integral-field spectrograph (Larkin et al.,
 117 2006) that operates in the z , J , H and K bands, with a nominal spectral
 118 resolution of $\lambda/\Delta\lambda = 3700$ (roughly $5\text{ \AA}$) and a diffraction-limited spatial
 119 resolution approaching 60 mas in K band provided by the adaptive optics
 120 system (Wizinowich et al., 2000). The instrument was mounted to the Nas-
 121 myth platform of the Keck II telescope over the timespan of this observing
 122 campaign and was moved to Keck I in 2012.

123 For most of the campaign we used the broadband Kbb ($1.965\text{--}2.381\text{ }\mu\text{m}$)
 124 filters at a 20 mas plate scale, which corresponds to a $0.32\times 1.28''$ field of view
 125 (FOV) with 1665 spectral channels. The observed diameter of Titan when
 126 including the atmosphere is $\sim 1''$, so 4 exposures were spatially mosaicked for
 127 full disk coverage. The OSIRIS reduction pipeline was used for the following
 128 standard processing steps: flat-fielding, sky subtraction, cosmic ray removal,
 129 channel level adjustment and the removal of crosstalk, before the extraction
 130 of spectra and assembly of a datacube. The reduction pipeline was recently
 131 described as part of the work by Iserlohe et al. (2013). Mosaicking via the
 132 OSIRIS pipeline was unreliable and was instead performed manually. Dat-
 133 acubes were collapsed over wavelength to create an image. The exposures
 134 near the center of Titan that sampled both the North and South limbs were
 135 used to fix the apparent diameter and a circle was drawn. The exposures
 136 near the East and West limbs were then offset to within a 20 mas pixel in
 137 order to produce a circular image of the disk. A gradual mechanical deteri-
 138 oration of the lenslet mask caused the leakage of light across channels and
 139 was manifest as artificially bright pixels at the edge of the FOV. These pixels
 140 were masked as necessary, however, since the magnitude of this artifact varies
 141 with wavelength, it was not possible to completely mask out all edge pixels
 142 without also excluding valid data. Therefore, artifacts at the edges of some
 143 images remain, manifested as bright vertical stripes in the mosaicked dat-
 144 acubes. To a lesser degree there are also faint vertical stripes near the center
 145 of the disk in some images. Later in the observing campaign we switched
 146 to narrow-band Kn2 ($2.036 - 2.141\text{ }\mu\text{m}$) and Kn3 ($2.121 - 2.229\text{ }\mu\text{m}$) filters,
 147 which at a 20 mas plate scale offer $0.90 \times 1.28''$ FOV that covers nearly the
 148 entire disk of Titan and eliminates the need for spatial mosaicking. The Kn2
 149 and Kn3 filters combine to have 854 spectral channels with slightly abridged

wavelength coverage relative to the Kbb filter. Depending on the angular diameter of Titan, and possible vignetting, we recorded ~ 1100 – 1500 spatial pixels on the disk, resulting in cubes of $\sim 2 \times 10^6$ datapoints in the Kbb filter and half that in the combination of Kn2 and Kn3 filters, for a total of 2.62×10^7 spatial-spectral datapoints in all of the observations.

On each night an A-type star with J , H , and K magnitudes from *2MASS* was used to correct for telluric absorption, instrument spectral response, and for photometry as described by Laver et al. (2007). A second-order photometric correction was performed using 113 IRTF spectra (Brown, M. E., *private communication*). The light curves for Titan at wavelengths sensitive to the surface (2.017 – $2.044 \mu\text{m}$) and stratosphere (2.216 – $2.266 \mu\text{m}$) were fit to the calibrated IRTF measurements and used to provide an offset and scale factor correction for the disk-integrated OSIRIS observations. The log of the observations is presented in Table 1, and disk-integrated spectra that characterize the datacubes are presented in Figure 1.

3. Radiative transfer model

Synthetic spectra are generated by defining 20 atmospheric layers, with properties that are determined primarily by measurements made with instruments on the *Huygens* probe. The layers have boundaries (levels) that are evenly spaced in pressure, with 10 levels above and 10 levels below 300 mbar (see Table 2 in Ádámkovics et al., 2016). The top of the atmosphere is set at zero optical depth. We use measurements of the temperature, pressure, methane abundance, and haze structure to determine the gas and scattering opacity in each layer. The *Huygens* temperature profile is used at all latitudes. CH_4 and CH_3D line opacities are from the HITRAN 2012 database (Rothman et al., 2013). The discrete-ordinate-method radiative transfer (DISORT; Stamnes et al., 1988) is implemented in Python (PyDISORT) and used to solve the radiative transfer through the model atmosphere and simulate the observed flux (Ádámkovics et al., 2016). We use the pseudo-spherical correction to plane-parallel geometry in DISORT to account for Titan’s thick atmosphere and limit our analysis to viewing geometries on Titan with solar zenith angles below 50° , where the correction is smaller than other sources of uncertainty.

Each (x, y) pixel in a datacube maps to a spectrum at a given latitude and longitude, with a corresponding incidence angle θ_0 , and emission angle, θ_e . Below we use the standard notation, $\mu_0 \equiv \cos \theta_0$ and $\mu_e \equiv \cos \theta_e$. Since $\mu_0 \approx$

Table 1: Log of Observations

Date (UTC)	Filter	^a Exp.	Airmass	^b Lat.	^b Lon.	^c Dia.	^d Phase	Cal. Star
2006-01-15 13:42	Kbb	400	1.17	-18.2	114.9	0.87	1.52	HIP 53721
2006-04-17 06:16	Kbb	360	1.04	-19.8	36.3	0.79	6.25	HIP 59271
2007-01-29 12:29	Kbb	300	1.03	-13.2	18.3	0.86	1.49	HIP 59271
2008-02-22 07:18	Kbb	360	1.58	-8.1	148.9	0.86	0.34	HIP 51213
2008-03-25 05:32	Kbb	360	1.38	-9.2	153.5	0.84	3.25	HIP 51213
2008-12-07 14:24	Kbb	360	1.26	-0.8	179.4	0.76	6.01	HIP 51213
2010-01-04 14:34	Kbb	360	1.11	5.0	40.1	0.77	5.89	HIP 62376
2010-01-29 13:51	Kbb	360	1.06	4.8	244.8	0.80	4.90	HIP 62376
2011-01-10 15:05	Kbb	360	1.14	10.3	124.8	0.75	5.89	HIP 69658
2011-03-15 13:36	Kbb	360	1.22	9.4	131.2	0.82	2.19	HIP 69658
2012-01-02 14:43	Kbb	360	1.51	14.8	254.2	0.71	5.55	HIP 69658
2013-04-06 14:26	Kn3	300	1.53	18.8	202.1	0.80	2.28	HIP 69974
2013-04-06 13:48	Kn2	300	1.35	18.8	201.5	0.80	2.28	HIP 69974
2015-07-25 07:14	Kn3	360	1.35	24.1	60.8	0.75	5.23	HIP 75939
2015-07-25 05:53	Kn2	360	1.26	24.1	59.5	0.75	5.23	HIP 75939
2015-07-28 06:04	Kn3	360	1.27	24.1	127.5	0.74	5.35	HIP 75939
2015-07-28 05:56	Kn2	360	1.26	24.1	127.3	0.74	5.35	HIP 75939

^aExposure time in seconds for individual FOV. Four exposures with the same integration time were used to obtain data over the entire disk, with the exception of 2008-12-07 when we used five exposures.

^bSub-observer latitude and West longitude at the start of the first exposure from *JPL Horizons* ephemerides.

^cApparent diameter of the solid surface. The observable atmosphere increases this size by $\sim 10\%$.

^dThe Sun-Target-Observer (S-T-O) phase angle.

Figure 1: Disk-integrated spectra for all targets. Offset for clarity. The observations from 2013 and 2015 used narrow-band Kn2 and Kn3 filters, which cover a smaller spectra bandpass but benefit from a larger FOV that does not require mosaicking.

μ_e for these low Sun-Titan-Observer phase angles, we discuss the path length through the atmosphere on Titan (or airmass) as generally characterized by an approximate $\mu \approx \mu_0$. A $2.018\ \mu\text{m}$ map of Titan (Barnes et al., 2007), updated with recent Cassini/VIMS measurements (Barnes, J. W., *private communication*) is to determine the surface albedo for each pixel. The limits in latitude and longitude for each pixel are calculated and the mean value of the rectangular footprint on the surface albedo map is used as input to the radiative transfer model, see Figure 2. We estimate a 10% systematic uncertainty in our surface reflectivity map, which is propagated through to our retrieved model parameters. With some constraint on the surface albedo, we can simulate a spectrum for a given viewing geometry that has only the properties of the haze as free parameters. The catalog of observations for the entire campaign is illustrated with parameters related to the viewing geometry and surface albedo in Figure 3.

3.1. Aerosol scattering sensitivity tests

Starting with a nominal model of the atmosphere, as measured by instruments on the *Huygens* probe, we perform a series of sensitivity tests to illustrate the spectral changes resulting from a variation in the haze optical depth. We make the assumption here that the DISR model at $2.0\ \mu\text{m}$ is a reasonable enough extrapolation of observations with maximum wavelength of $1.6\ \mu\text{m}$ to be a valid starting point for our models. The sensitivity tests are used to inform the spectral fitting algorithm described below in Section 3.2, and to illustrate the limitations and uncertainties in the analysis. The tests are performed by adding a perturbation of $\delta\tau_H$ to the nominal haze opacity in a given layer. The difference between the simulated spectrum with the perturbation and the nominal spectrum is the spectral response, $\Delta I/F(\lambda)$, at each wavelength λ . The spectral response depends on the altitude of the haze perturbation, Figure 4. For example, increased haze opacity in the 53.5 km altitude layer of the model increases the reflectivity at all wavelengths (purple), whereas adding haze near the surface (red) does not change the spectrum at wavelengths longer than $2.15\ \mu\text{m}$. The wavelength region near $2.12\ \mu\text{m}$, which is generally thought to be diagnostic of clouds in the 20 – 40 km altitude range (e.g., Brown et al., 2002; Griffith et al., 2009; Ádámkovics et al., 2010), is also sensitive to haze opacity at other altitudes. This degeneracy means that a unique solution to the vertical structure of haze is not retrieved from an individual spectrum. The degeneracy is also part of the motivation for analyses with simplified scalings of the haze opac-

Figure 2: Surface reflectivity reprojection. An example of the surface albedo determination for each pixel in an observation is illustrated for the observations on 2008-02-22. A $2.018\ \mu\text{m}$ VIMS surface albedo map (top; updated from Barnes et al., 2007) is used to determine the mean surface albedo in the approximate footprint (orange squares) covered by a pixel. These values are used to generate a synthetic surface albedo for the observed disk (bottom left), which can be compared to observations at wavelengths that are sensitive to the surface (bottom right; $2.017\text{--}2.044\ \mu\text{m}$).

Figure 3: Catalog of observations. For each night of observations, two characteristic images that are sensitive to the surface (2.017–2.044 μm) and stratosphere (2.215–2.265 μm) are shown in grayscale. The color panels illustrate the latitude and longitude, and the right-most panel in each set (grayscale) is the reprojected surface albedo map that corresponds to the viewing geometry. Mosaicking artifacts can be seen to be particularly prominent in 2010 and 2011, before being nearly completely eliminated in the observations between 2013 and 2015. Seams from the surface albedo map also show up in some of the reprojections. Uncertainty due to the surface albedo map is discussed in the text.

Figure 3: (continued)

Figure 3: (continued)

ity above and below a particular altitude in the atmosphere, e.g., Karkoschka (2016).

The altitude sensitivity tests can be used to select an altitude boundary in the model for the prescribed changes to the haze vertical structure. We define two independent regions of “upper” and “lower” hazes separated at the altitude z_{crit} . The center of the K-band window near $2.16\ \mu\text{m}$ is the wavelength cutoff for sensitivity to hazes above and below 18.8 km (green dashed spectra in Figure 4). Adding haze opacity to the model below ~ 20 km altitude does not impact the spectral fitting for $\lambda > 2.16\ \mu\text{m}$, since $\Delta I/F(\lambda) = 0$. This means that roughly one half of spectral channels in the observations are sensitive to only the haze above ~ 20 km, while the other half are sensitive to both the hazes above and below. The model atmosphere is also setup such that roughly half the model layers are above and below this altitude (see Table 2 in Ádámkovics et al., 2016). The altitude boundary at 20 km altitude facilitates fitting spectra in the $2\ \mu\text{m}$ window but otherwise has no physical motivation.

Although the haze opacity is optically thin in each layer of the model, we test for non-linearity in the spectral response to $\delta\tau_H$ by comparing $\Delta I/F$ for values of $\delta\tau_H = 0.1$ and 0.3 . The similarity of the panels on right and left sides of Figure 5 demonstrates that for an optically thin increase in haze opacity the magnitude of spectral response is proportional to $\delta\tau_H$. Indeed, there are no significant variations in the shape of the spectral response for the two cases shown in Figure 5. However, this is only demonstrated for an optically thin example, and would not necessarily remain true for optically thick increases in haze opacity.

Although the shape of the spectral response does not depend on $\delta\tau_H$, it does depend on the slant-path through the atmosphere, as determined by the viewing geometry. Adding the same opacity at larger airmass (smaller μ) means that the same opacity should result in a larger $\Delta I/F$ at some wavelengths. We confirm and quantify this picture by comparing the spectral response to $\delta\tau_H = 0.1$ for two viewing geometries (Figure 6). Comparing the left and right columns in Figure 6 confirms that the spectral response is larger for a given opacity increase in the higher-altitude regions of the atmosphere. On the other hand, the difference in the spectral responses at wavelengths sensitive to the lower atmosphere is smaller.

Figure 4: Altitude sensitivity of haze scattering. The nominal synthetic reflectivity spectrum (top; black) is compared to spectra where an increase in the haze opacity of $\delta\tau_H = 0.1$ is added to 10 different altitude layers in the model (color spectra). The altitude at the bottom of the model layer z_{min} is specified in the legend. The spectral response $\Delta I/F(\lambda)$ (bottom) is the difference between the model reflectivity for two simulations; one with the added haze opacity $\delta\tau_H$ in one layer of model, as compared to the nominal reference model, for a given wavelength, λ .

Figure 5: Optical depth sensitivity of haze scattering. The spectral response plotted in the bottom panel of Figure 4 is shown here, but in this case normalized by the magnitude of the optical depth increase, $\delta\tau_H = 0.1$ (top left). The spectral response for a $3\times$ larger increase of $\delta\tau_H$ is also shown (top right). The similarity of the expanded spectral regions in the bottom row illustrates the linear spectral response to an optically thin increase in haze scattering at various altitudes.

Figure 6: Airmass sensitivity of haze scattering. The spectral response $\Delta I/F(\lambda)$ due to an increase in the haze opacity by $\delta\tau_H = 0.1$ at various altitudes is shown by the colored spectra for an airmass of $\mu = 0.3$ (left) and compared with the corresponding spectra for and airmass of $\mu = 0.9$ (right). In the case of the larger airmass (smaller μ), $\Delta I/F$ can be up $\sim 20\%$ larger or smaller, depending on the altitude of the haze increases. For example, increasing the haze in the 53.5 km layer of the nominal model at low airmass ($\mu = 0.9$ right) corresponds to $\Delta I/F(2.19) = 0.007$, whereas at larger airmass ($\mu = 0.3$; left), there is a larger $\Delta I/F(2.19) \sim 0.009$. The opposite trend occurs for a haze increase in the 28.4 km layer of the model near $2.13 \mu\text{m}$, where the response is smaller at large airmass. The illustrative case of $\mu = 0.3$ is an extreme example used to bracket our modeling uncertainty, and we do not consider $\mu < 0.68$ when fitting observations.

3.2. Spectrum fitting

Fitting each individual spectrum in the observations with the PyDISORT radiative transfer model is computationally expensive. If each of the 26.2 million (spatial and spectral) datapoints from the observations could be fit in 1 s, it would take 10 months to fit all of the observations. However, the computing time for running a single monochromatic channel with PyDISORT on a standard 24-core workstation is a few seconds, and it takes (at least) a few iterations of the calculation for a fitting algorithm to converge. This makes fitting of the entire dataset of Keck OSIRIS observations unfeasible with a standard workstation.

In order to minimize the amount of data that is simulated, and reduce the number of radiative transfer calculations that are required, we (1) average the spectra in latitude bands on Titan, thereby integrating across longitudes that are assumed to have similar haze properties, and (2) reduce the total number of channels in the spectra that are fit during the optimization, by assuming that multiple channels redundantly sample the same altitude regions of the atmosphere. The fit is confirmed with a full spectrum calculation after the parameter optimization is complete. These two approximations are described in greater detail below.

Latitude band integration: The hazes on Titan can vary significantly with latitude, but much less so with longitude (Smith et al., 1981, 1982; Lorenz et al., 2001; Rannou et al., 2000; de Pater et al., 2006; Anderson et al., 2008; Lorenz et al., 2010; Penteado et al., 2010; de Kok et al., 2010; Rannou et al., 2010; Larson et al., 2015). We take advantage of the symmetry over longitude to combine spectra and reduce the number of calculations needed, while also improving the signal to noise ratio of the observation being simulated. However, there is a significant variation in the spectral response with airmass (Figure 6), and spectra from various paths through the atmosphere are obtained from within a latitude region. Therefore, we evaluate the number of spectra that can be combined without introducing significantly increased uncertainty in the combined spectra due to variation in the viewing geometry. For example, in the observations on 2008-02-22 there are 40 spectra in the latitude region bounded by 26.9° S and 29.8° S. In the Supplementary Material we illustrate that the systematic uncertainty introduced by combining the 20 spectra at lowest airmass, with $\mu > 0.8$, is negligible.

The integration over latitude band regions is performed over the entire disk. Starting from the south pole, the latitude limit that encompasses 40 pixels is determined (e.g., from 90° S to 66.1° S on 2008-02-22), and the band

296 integration is performed by considering the 20 spectra with the smallest
 297 airmass. The excluded pixels occur near the East and West limb, with the
 298 low airmass pixels coming from near the center of the disk. The process is
 299 repeated northward. The latitude band regions near the north and south limb
 300 have a low median μ , which corresponds to a high airmass that is challenging
 301 to treat in the pseudo-spherical radiative transfer model. In the subsequent
 302 analysis of retrieved haze parameters, the latitude band regions that do not
 303 have a median $\mu > 0.7$ are ignored.

304 The largest airmass that is considered in this work is 1.48, correspond-
 305 ing to a viewing geometry with $\mu = 0.68$ and solar zenith angle of 48° .
 306 In general, most datasets have a maximum airmass of ~ 1.3 , such that the
 307 pseudo-spherical correction is an applicable correction to the radiative trans-
 308 fer calculation.

309 *Spectral region selection:* Wavelengths in the $2\mu\text{m}$ window are sensitive
 310 to various altitudes throughout the atmosphere and surface, e.g., see con-
 311 tribution functions in Figure 6 in Ádámkovics et al. (2006). For the $2\mu\text{m}$
 312 window in general, the fainter regions of low I/F (with large gas opacity)
 313 are sensitive to higher in the atmosphere, whereas bright regions (with lower
 314 gas opacity) are sensitive to the surface. It can be approximated that regions
 315 of similar I/F are roughly sensitive to similar altitude regions. Due to this
 316 redundancy, we do not fit all 1665 spectral channels in the K-band. Instead,
 317 we pick seven wavelength windows listed in Table 2 that sample altitudes
 318 throughout the atmosphere, using a total of 104 OSIRIS spectral channels.
 319 By considering this subsample of 104 channels from the integrated spectra
 320 for ~ 29 latitude bands, we reduce the total number datapoints being fit by
 321 a factor of ~ 800 for each observation.

322 *Parameter optimization:* We fit each spectrum with a model that in-
 323 cludes two free parameters; one parameter f_S scales the haze opacity above
 324 a critical altitude $z_{crit} = 20\text{ km}$ and a second parameter f_T scales the haze
 325 opacity below this altitude. Since the surface albedo is determined from
 326 the surface reflectivity map, and we assume the temperature and methane
 327 abundance profiles measured by instruments on *Huygens*, there are no other
 328 unconstrained parameters. Starting from the standard reference spectrum,
 329 a Levenberg-Marquardt (LM) optimization is used to simultaneously fit f_S
 330 and f_T . For simplicity in our approximate approach, we consider z_{crit} to be
 331 fixed. A higher (lower) value for z_{crit} means that the total column of haze
 332 opacity being scaled by each parameter is different and would result in a
 333 systematically smaller (larger) value retrieved value for f_T .

Table 2: Spectral Regions for RT Model Optimization

Wavelength (μm)	Channels
2.042 – 2.045	12
2.072 – 2.074	8
2.092 – 2.094	8
2.101 – 2.105	16
2.132 – 2.135	12
2.141 – 2.145	16
2.182 – 2.186	16
2.200 – 2.204	16

Although the shorter wavelengths in the K-band window are sensitive to changes in both parameters, the altitude sensitivity test (above) suggests that a two-parameter optimization of the model with a boundary at 20 km is not degenerate, since the longer wavelengths can be used to independently constrain f_S . In practice we use `scipy.optimize.curve_fit` with a function that is a wrapper to a PyDISORT calculation for the 104 channels in the seven wavelength windows. The LM optimization usually converges with ~ 10 spectral calculations. Once f_S and f_T have been optimized for the spectrum from a latitude band region, a full spectrum over the 1665 channels is calculated for direct comparison and validation.

4. Results

The best fit spectra for a range of latitude band regions are compared with the observations from 2008-02-22 in Figure 7. The optimized values for f_S and f_T , along with the mean surface albedo in the latitude band, $\bar{A}_S$, are shown for each panel. The models generally reproduce the observations for the spectra at $\lambda > 2.14 \mu\text{m}$. This is primarily because these wavelengths are insensitive to the surface albedo and are sufficiently modeled with a single-parameter scaling of the haze above 20 km. It is important to note that as f_S varies the haze in the uppermost layers of the model, which $\lambda > 2.14 \mu\text{m}$ are most sensitive to, this spectral region and the retrieved f_S is not particularly sensitive to z_{crit} . However, as described above, the retrieved value of f_T does depend on the z_{crit} . The largest model error at $\lambda > 2.14 \mu\text{m}$ occurs

in the two latitude band regions nearest the north and south limb, where the simulated spectra overestimate the I/F near $2.20\ \mu\text{m}$. The overestimate in the spectral region with the smallest I/F , sensitive to the highest altitudes in the atmosphere, demonstrates the limitations of the pseudo-spherical correction in the plane parallel model. However, since the latitude regions near the limb have a broad range of airmass, we’ve already excluded them from the analysis (Section 3.2) and do not consider the retrieved values of f_T and f_S in the analysis below.

At shorter wavelengths ($\lambda < 2.14\ \mu\text{m}$), the most significant discrepancy between the simulated spectra and observations occurs near $2.09\ \mu\text{m}$ (Figure 7). This wavelength region is sensitive to haze variation throughout the atmosphere (e.g., Figures 4 and 6), and therefore decreasing f_T in order to decrease I/F near $2.09\ \mu\text{m}$ would also decrease I/F near $2.05\ \mu\text{m}$ and degrade the fit. The limitation in the spectral fitting arises from only using two parameters to model the variation in the haze distribution. Using additional free parameters would improve the fit, but would lead to degeneracies among the haze scaling parameters. Additional free parameters would also significantly increase the calculation time of the LM parameter optimization.

One possible source of error is a systematically underestimated surface albedo from the brightest regions of the VIMS albedo map. Increasing the surface albedo that is used in the simulations would result in a fit with a decreased f_T . Increasing the surface albedo and decreasing f_T would improve the fit to the observations at both $2.05\ \mu\text{m}$ and $2.09\ \mu\text{m}$. Alternatively, the gas opacity in this wavelength region may be underestimated in the model, or the haze single scattering albedo in this wavelength region may be too high. Both the gas opacity and the haze scattering albedo are assumed to be uniform at all locations.

The retrievals of f_S and f_T that have been illustrated for 2008-02-22 are performed for each of the observations. The meridional trend in haze, as represented by the best fit parameter f_S , is characterized by a seasonally variable linear gradient in haze toward the North pole, together with a seasonally variable decrease in opacity poleward of 20° (Figure 8). The magnitude of the linear gradient toward the North increases with time over the timespan of the observations from 2006 through 2013, before starting to decrease. The overall trend is complicated by the poleward decrease in opacity. While the decrease near the limb raises suspicions of a modeling or retrieval artifact, there are a few considerations that suggest the decreases may be real: (1) in nearly all of the observations, with the exception of 2010-01-29, the decrease

Figure 7: Two-parameter model optimization for f_S and f_T . The best fit models (green spectra) and the observations (black spectra) that are integrated over a particular latitude band (green pixels in inset image) are illustrated for each latitude band (panels). The observations from 2008-02-22 are shown. For each panel, the latitude range for the observed spectrum is given in parenthesis. The best fit haze scale factors; f_S above 20km and f_T below 20km, are listed in each panel with the median surface albedo in the region $\bar{A}_S$. The inset image of is integrated over 2.21 — 2.26 μm .

394 is either absent or occurs asymmetrically toward one limb. A geometric ef-
 395 fect whereby the retrieval systematically underestimates f_S near the limb
 396 is anticipated to result in symmetric underestimates at both the North and
 397 South edges of the disk. (2) The decrease in f_S toward the limb varies among
 398 the observations, sometimes occurring in the South (2011) and sometimes
 399 occurring in the North (2006). A geometric artifact, even if asymmetric for
 400 some reason, would not be seasonally variable. (3) the latitude regions that
 401 are closest to the limb, where retrieval artifacts due to model limitations at
 402 high airmass are most significant, have been excluded and are not plotted in
 403 Figure 8. The poleward decrease in f_S may indeed be an indication of the
 404 formation of features like the polar collar (Smith et al., 1981; Lorenz et al.,
 405 2001; de Pater et al., 2006).

406 The plot of the f_T vs latitude (Figure 9) demonstrates a general corre-
 407 lation between the surface albedo and the lower atmospheric haze. Along
 408 with this correlation, there are noteworthy changes in the tropospheric haze.
 409 For example, In 2006 there was a ‘bulge’ in southern tropospheric haze, with
 410 the peak of the bulge shifting southward by 20° latitude over three months.
 411 Features in these plots are likely due to uncertainties in the absolute value
 412 of the surface albedo map. Since it is difficult to remove a uniform contri-
 413 bution from atmospheric scattering to the entire map, there remains some
 414 uncertainty in the surface reflectivity. If the surface reflectivity is systemati-
 415 cally (over-) underestimated, then the optimized value for f_T will be (under-)
 416 overestimated. In the case of the 2013 observations, the limits of the absolute
 417 flux calibration are reached due to cloudy observing conditions. The best fit
 418 parameter for the f_T is the complete removal of haze below 20 km, however,
 419 the observations are consistent with uniform tropospheric haze.

420 There are large negative excursions from the otherwise relatively smooth
 421 meridional trends in the retrieved values for f_T (Figure 9) on 2008-02-22,
 422 2008-03-25, 2011-01-10, and 2011-03-15. On each of the dates, the small val-
 423 ues in f_T occur between the equator and 10° S latitude. The sub-longitude
 424 on these dates are 149° W, 153.5° W, 125° W, and 131° W. This suggests
 425 that there is a geographic area constrained in both latitude and longitude
 426 where the haze scattering opacity needs to be completely removed in order
 427 to fit the observations. On the other hand, it has been suggested that the
 428 methane opacity in this region, corresponding to the Xanadu region, might
 429 be preferentially enhanced (Ádámkovics et al., 2007, 2009). If this is the
 430 case, then the variation in methane opacity would allow for more uniform
 431 variation in the haze, as is seen at other longitudes. In other words, by invok-

ing an increased methane opacity near Xanadu, the fitting algorithm would
 not converge on a decrease f_T , and the smoother latitudinal variation that
 is observed on other dates would occur in all cases. An alternate explanation
 is that the brightness of the surface albedo near Xanadu in the VIMS
 maps is overestimated. Overestimating the surface albedo would result in
 artificially bright simulated spectra, which would result in the optimization
 returning small (or zero) values for f_T . However, this somewhat contradicts
 the interpretation of the shape of the simulated spectra, in particular the
 (under)overestimate of I/F at $(2.04)2.09\ \mu\text{m}$, which could be explained by a
 brighter surface albedo map. An interpretation that we cannot rule out is
 that perhaps both the overall reflectivity of the VIMS surface albedo map is
 too low while the reflectivity of Xanadu specifically is too high.

5. Discussion

The hazes on Titan have been studied with extensive observations (e.g.,
 Smith et al., 1981, 1982; Coustenis and Bezard, 1995; Hutzell et al., 1996;
 Lorenz et al., 2001; Gibbard et al., 1999; Rannou et al., 2000; Tomasko et al.,
 2005; Ádámkovics et al., 2006; Anderson et al., 2008; Rannou et al., 2010;
 Lorenz et al., 2010), and models of circulation predict that the winds in
 the atmosphere seasonally redistribute the haze (e.g., Rannou et al., 2002;
 Larson et al., 2015). Recently presented analysis of *HST-STIS* observations
 by Karkoschka (2016) suggest that the observations of haze can be well-
 approximated by a sinusoidal variation in haze components above and below
 $\sim 100\text{ km}$, with a phase lag between the two components and a meridional
 dependence in the magnitude of the sinusoid variation. Our results suggest
 that such a model may be an oversimplification of the haze variation. We
 observe meridional structure in the haze that is often characterized significant
 asymmetries and abrupt variations in the opacity of haze with latitude.

Penteado et al. (2010) studied VIMS datasets covering the timespan of
 July 2004 - March 2007 simulating spectra in the $0.4 - 1.6\ \mu\text{m}$ range. The
 shorter wavelengths (relative to this study at $2\ \mu\text{m}$) are sensitive to higher
 altitudes in the atmosphere. Penteado et al. (2010) fit their VIMS spectra
 with both haze opacity and scattering albedo variation above 80 km . They
 find that the haze opacity above 80 km increases from ~ 0.9 near 60°S to
 1.1 at 40°N and that the scattering albedo decreases by $\sim 20\%$ toward the
 North over this region. We have temporal overlap with three of our datasets,
 (2006-01-15, 2006-04-17, and 2007-01-29), and can make a comparison to

Figure 8: Meridional variation in the haze above 20 km altitude. LM optimization is used to determine the scale factor f_S , for each latitude region in each of the observations (panels). The date is indicated in the top left and the sub-observer West longitude indicated in the bottom right of each panel.

Figure 9: Meridional variation in the haze below 20 km altitude. LM optimization is used to determine the scale factor f_T , for each of the observations (panels). The date is indicated in the top left and the sub-observer West longitude indicated in the bottom right of each panel.

468 their results, where both studies consider the case of uniform methane and
469 evaluate the variation in haze.

470 Comparing Figure 16 of Penteado et al. (2010) (covering 2004-2007) to
471 Figure 8 of this work (for individual dates), it can be noted that the multi-
472 plicative scale factors in both studies are generally in agreement: ~ 0.9 near
473 50°S and increase northward to a factor of ~ 1.1 near 20°N . An interesting
474 bump in the haze opacity near 60°S in the Penteado et al. (2010) is also
475 reproduced in our observations in 2006, but is gone by 2007. And important
476 difference is the drop in haze opacity that we observe north of $\sim 10^\circ\text{N}$. Some
477 amount of the discrepancy may be due our assumption of uniform single-
478 scattering albedo. Penteado et al. (2010) implement a decrease in scattering
479 albedo toward the north that could account for the some of the decrease in
480 reflected flux that we simulate with decreased haze opacity. On the other
481 hand, the decrease in haze opacity with latitude that we retrieve is steeper
482 than the decrease in scattering albedo retrieved by Penteado et al. (2010).

483 Our observations and analysis are complementary to recent *HST-STIS*
484 analysis (Karkoschka, 2016) in many ways, including the wavelength region
485 and resolution, as well as the timing and time-span of the observations. Ta-
486 ble 3 lists the characteristics of the datasets presented here and by Karkoschka
487 (2016). In addition to the characteristics of the datasets, the method of anal-
488 ysis is also complementary. The large number of datapoints in both studies,
489 and the calculation time required for generating simulated spectra, means
490 that some method of reducing the total number of datapoints need to be
491 invoked. While Karkoschka (2016) use a principle components analysis to
492 identify the relevant features of the dataset, we use a two-parameter LM
493 optimization to fit the spectra. Rather than avoiding the spectral bands
494 that are most sensitive to the surface, we include them and model the entire
495 K-band.

496 A quantitative comparison to the work of Karkoschka (2016) should not
497 simply compare haze scale factors because of different choices how the hazes
498 are varied in the radiative transfer models. One important consideration is
499 the altitude region considered; above and below 20 km here, vs above 150 km
500 and below 80 km in Karkoschka (2016). This is in addition to differences
501 in the nominal haze model as well as wavelength sensitivities. Rather than
502 compare derived haze opacity scale factors, we take a closer look at the
503 corresponding $2\ \mu\text{m}$ optical depth variation. The nominal $2\ \mu\text{m}$ optical depths
504 for the two models that are based on the DISR observations are shown in
505 Figure 10. There is a haze extinction optical depth of, $\tau_H=0.13$ above 150 km

Table 3: Comparison with *HST* analysis

	This work	Karkoschka (2016)
Instrument	Keck OSIRIS	<i>HST-STIS</i>
Spectrometer	IFU	slit
Wavelength	1.965 – 2.381 μm	0.530 – 1.020 μm
Plate scale	20 mas/pixel	50 mas/pixel
$R = \lambda/\delta\lambda$	3,700	1,500
Obs. Start	2006-01-15	1997-11-03
Obs. End	2015-07-28	2004-01-03
Datapoints	26.2×10^6	$\sim 10^6$
Number of Obs.	14	5
Model		
Spectral bands	104	25
Latitude regions	29	10
Locations ¹	20	2

¹ The number of “center-to-limb” locations, or viewing geometries used to model each latitude region.

Figure 10: DISR haze models of the cumulative haze optical depth from the top of the atmosphere as reported by Tomasko et al. (2008), used as the nominal case for the radiative transfer models here, and the recent revision by Doose et al. (2016), which is the nominal model used by Karkoschka (2016).

506 and $\tau_H=0.37$ from 80 km to the surface in the Doose et al. (2016) model
507 DISR haze. The two haze scale factors, and their seasonal variation, that
508 are reported by Karkoschka (2016) can be converted variation in τ_H at $2\mu\text{m}$.
509 These curves are illustrated in Figure 11 (refer to the right axis of Figure 15
510 in Karkoschka (2016) for comparison).

511 In our nominal models, following Tomasko et al. (2008), we have $\tau_H=0.26$
512 below 20 km which we use to convert our scale factor f_H to τ_H for comparison.
513 Figure 10 shows that there is about $\tau_H=0.63$ above 20 km in the Tomasko
514 et al. (2008) model. Using this scale factor would result in factors of a few
515 to several larger variation in optical depth than in reported by Karkoschka
516 (2016). This is likely due to the altitude sensitivity of our radiative transfer

517 model. Although we vary the column of haze above 20 km in our model to
 518 determine f_S , the bulk of the spectral response in our model is from the
 519 upper-most layers of the model. The bottom level of the first layer of the
 520 model has a nominal optical depth of 0.11 in the Tomasko et al. (2008)
 521 model, which we used to scale our f_S to τ_H for comparison with the results
 522 of Karkoschka (2016). Figure 11 shows a comparison of the optical depth
 523 variation at low and high altitudes for six latitude regions compared to the
 524 optical depth variation predicted by the seasonal model of Karkoschka (2016).
 525 While there is general agreement in the magnitude of change and some trends
 526 with latitude, there is significantly larger variation within individual latitude
 527 bands than predicted from the seasonal model.

528 Although significant decreases in stratospheric haze are consistent with
 529 spacecraft measurements, e.g., the dark band observed by Voyager (Smith
 530 et al., 1981), the decreases in f_S near the poles are only faintly visible, if at
 531 all, in the images presented here that are sensitive to the stratosphere (e.g.,
 532 on 2006-01-15 and 2010-01-29, in Figure 3). At wavelengths with significant
 533 methane absorption the atmosphere becomes optically thick for a column of
 534 haze scattering that is optically thin. For increasing airmass near the limb,
 535 the increased slant path through low density gas corresponds to larger scat-
 536 tering optical depths, and brighter observed flux. Although we exclude pixels
 537 close to the limb, at $\mu < 0.68$, there may be effects of limb brightening that
 538 are exacerbated due to the observational point spread function (PSF). The
 539 proximity of the decrease in haze scattering opacity next to the significant
 540 limb brightening may decrease the contrast in the images. Artifacts in the
 541 data are also a concern and may obscure the features in the images sensitive
 542 to the upper haze layers. The mechanical lenslet mask for OSIRIS degraded
 543 in (re)positioning accuracy over time, causing scattered light from neigh-
 544 boring lenslets to contaminate the edge of the FOV, with the lenslet mask
 545 breaking completely in 2011. These lenslet masking artifacts near the edge
 546 of the OSIRIS FOV cause the vertical striping that is either faintly or, as in
 547 the case of observations on 2011-01-10, dramatically visible.

548 One of the primary concerns in interpreting the structure in upper haze
 549 variation seen in Figure 8 is possible artifacts in the retrieval of f_S near the
 550 limb. Along with the arguments described above to support the validity of
 551 these features, we can also compare our results to similar analysis of the haze
 552 opacity variation using *Cassini* VIMS. Rannou et al. (2010) use a radiative
 553 transfer model with a correction for spherical geometry (SPDISORT; Mayer
 554 and Kylling, 2005) to interpret a full disk VIMS image (13° phase angle), and

Figure 11: Seasonal variation in hazes opacity from the model of Karkoschka (2016) (curves, reproducing their line styles), with some latitudes highlighted in color for comparison; 45°N (blue), 25°N (green), 5°N (cyan), 5°S (yellow), 25°S (orange), 45°S (red). Datapoints retrieved from corresponding latitudes over the timespan of the observations are plotted in matching colors.

555 find a linear increase in haze from 80° S to the Equator, where there is a
556 sharp transition, and then uniform haze to $\sim 20^\circ$ N, and then a decrease in
557 opacity to 60° N (their Figure 3). Their analysis of an image of the North
558 polar region (at 113° phase angle) from 2006-12-22 confirms the decrease in
559 haze toward the North polar region. Our observations from 2006-04-17 and
560 2007-01-29 are closest in time; the former shows a discontinuity in the haze
561 in the tropics (although not at the Equator as in 2004), while the latter has a
562 corresponding decrease in haze from 20° N to the pole. The consistent inter-
563 pretation between the different observations, made using different radiative
564 transfer models, suggests that the poleward decreases in f_S reported here
565 (Figure 8) are real structure.

566 At 40° S we observe a decrease in f_S from 2006 through 2011, after which
567 this latitude is too close to the limb for accurate retrieval. Similarly, at 40° N,
568 we see an increase in haze from 2009 through 2013. At mid-latitudes, and in
569 the tropics, there’s no apparent trend with season. A transition near 20° S
570 observed in 2006 and 2008, but is not seen in 2007. In 2011, there a sharp
571 drop in haze from the Equator southward through 20° S, without a transition.
572 Linear gradients in the absence of either significant haze decreases near the
573 limb or discontinuities are only observed in 2013 and 2015. The lack of
574 simple trends in f_S with latitude and season haze suggests that a stochastic
575 component of the haze variation could be important.

576 The haze in the lower atmosphere (Figure 9) is even more challenging
577 to interpret in terms of a simple model of seasonal haze variation. In the
578 lower atmosphere there are various bumps and wiggles atop possible linear
579 meridional gradients in f_T . One of the interesting features in the retrievals of
580 haze below 20 km is the decrease in f_T between the $\sim 20^\circ$ S and the Equator,
581 on 2008-02-22, 2008-03-25, 2011-01-10, and 2011-03-15. On each of these
582 dates, the sub-observer latitude (Table 1 and bottom right in each panel
583 in Figure 9), is within 30° of 125° W, which is the bright region of Xanadu.
584 The results here would suggest that there is a persistent hole in the haze
585 below 20 km near Xanadu, which seems an unlikely interpretation of the
586 observations. We’ve previously argued for an enhancement in methane opac-
587 ity near Xanadu, with the interpretation of increased methane condensation
588 (Ádámkovics et al., 2007, 2009). Here, since we assume a uniform methane
589 opacity, we find that the parameter optimization for fitting the haze must
590 decreased the haze to zero in order to fit the spectra. Decreasing the haze
591 opacity below 20 km decreases the scattering and lowers I/F , resulting a
592 spectral response that is similar to an increase in methane absorption. The

spectra over Xanadu are systematically different than from other locations on Titan, and it seems that a spatially localized increase in the methane opacity is more likely than a localized decrease in haze scattering. For example, clouds are supposedly indications of (relatively) small spatial scale variations in methane. On the other hand, variations in the haze seem to occur on a much larger scale (with the exception of near the poles) and are generally (if not always) uniform along latitude bands. If an increase in methane opacity is indeed the correct, it still remains unclear whether or not methane is in the form of an additional absorber in the atmosphere, a coating on haze particles, or a layer on the surface. An additional consideration for the possibility of increased methane columns over Xanadu is the regional topography. Cassini RADAR SARTopo and altimetry has shown the region of Xanadu to be topographically low region (Lorenz et al., 2013). There is roughly 3 km amagat of CH_4 in the nominal model (or a column of $\sim 10^{25} \text{ cm}^{-2}$), which would be increased by 10% if there were an additional km of path-length appended to the bottom of the atmosphere. An increase in CH_4 of this magnitude is not sufficient to account for the decrease in flux that is commensurate with the decrease is f_T that we observe.

In the radiative transfer models presented here, methane opacities in spectral bands are implemented using correlated k-coefficients (Goody et al., 1989) rather than mean opacities per spectral bands. Methane line lists (e.g., HITRAN2012) go as short as $\sim 1.08 \mu\text{m}$, and do not cover the spectral regions in the *HST* analysis, therefore mean opacities must be used for that work. However, averaging over $5\text{\AA}$ in the near-IR corresponds to hundred of transitions that are not resolved. Within an individual spectral band, there will be regions of high and low methane opacity, which are sensitive to the surface and upper atmosphere. Averaging opacities to a mean value results in an intermediate opacity with sensitivity to an intermediate altitude range. Therefore, a mean opacity for a spectral band that includes strong methane opacity variation may have a larger uncertainty in the altitude sensitivity. On the other hand, the simplistic approximation of linear scaling of haze above and below a single altitude boundary likely outweighs the uncertainties due to how the gas opacities are implemented in the radiative transfer.

6. Conclusions

We summarize and make the following conclusions from this analysis:

- 628 • Seasonal trends in the haze above and below 20 km on Titan are re-
629 trieved from 9.5 years of Keck OSIRIS observations in the K-band using
630 a radiative transfer model and LM optimization.
- 631 • The Northward linear gradient in haze above 20 km is punctuated by
632 significant seasonal decreases near the poles.
- 633 • The haze below 20 km is not well described by a simple seasonal model
634 with stochastic variations playing a potentially significant role.
- 635 • Retrievals of the haze near Xanadu are consistent with the suggestion
636 that the methane opacity is underestimated there.

637 The archive of Cassini VIMS observations over this period provide a tremen-
638 dous resource for comparison and validation of the variation in haze. Ex-
639 tending the approach presented here to systematically model the hazes for a
640 significant set of VIMS observations would be the next step in determining
641 the relative importance of seasonal vs. stochastic variation in the hazes on
642 Titan.

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