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A model intercomparison of Titan’s climate and low-latitude environment

Juan M. Lora^{a,b}, Tetsuya Tokano^c, Jan Vatant d’Ollone^d,
Sébastien Lebonnois^d, Ralph D. Lorenz^e

^a*Department of Earth, Planetary and Space Sciences, University of California, Los Angeles, California, USA*

^b*Department of Geology and Geophysics, Yale University, New Haven, Connecticut, USA*

^c*Institut für Geophysik und Meteorologie, Universität zu Köln, Köln, Germany*

^d*Laboratoire de Météorologie Dynamique, IPSL, CNRS/UPMC, Paris, France*

^e*Johns Hopkins Applied Physics Laboratory, Laurel, Maryland, USA*

Abstract

Cassini-Huygens provided a wealth of data with which to constrain numerical models of Titan. Such models have been employed over the last decade to investigate various aspects of Titan’s atmosphere and climate, and several three-dimensional general circulation models (GCMs) now exist that simulate Titan with a high degree of fidelity. However, substantial uncertainties persist, and at the same time no dedicated intercomparisons have assessed the degree to which these models agree with each other or the observations. To address this gap, and motivated by the proposed Dragonfly Titan lander mission, we directly compare three Titan GCMs to each other and to *in situ* observations, and also provide multi-model expectations for the low-latitude environment during the early northern winter season. Globally, the models qualitatively agree in their representation of the atmospheric circulation and structure, though one model severely underestimates

Email address: juan.lora@yale.edu (Juan M. Lora)

zonal winds and latitudinal temperature gradients. We find that, at low latitudes, simulated and observed atmospheric temperatures closely agree in all cases, while the measured winds above the boundary layer are only quantitatively matched by one model. Nevertheless, the models simulate similar near-surface winds, and all indicate these are weak. Likewise, temperatures and methane content at low latitudes are similar between models, with some differences that are largely attributable to modeling assumptions. All models predict environments that closely resemble that encountered by the Huygens probe, including little or no precipitation at low latitudes during northern winter. The most significant differences concern the methane cycle, though the models are least comparable in this area and substantial uncertainties remain. We suggest that, while the overall low-latitude environment on Titan at this season is now fairly well constrained, future *in situ* measurements and monitoring will transform our understanding of regional and temporal variability, atmosphere-surface coupling, Titan's methane cycle, and modeling thereof.

Keywords: Titan, Climate, Atmospheres, Meteorology

1. Introduction

2 The Cassini-Huygens mission transformed our understanding of Titan,
3 clarifying many aspects of its climate system but also leaving tantalizing
4 unanswered questions. Remote sensing observations, building on prior and
5 ongoing ground-based efforts, revealed a perplexing scarcity of tropospheric
6 methane clouds, but also demonstrated the existence of strong seasonal evo-
7 lution and the occurrence of precipitation (Turtle et al., 2011a,c; Rodriguez

8 et al., 2011; Turtle et al., 2018b). Lakes and seas, as well as flooded canyons,
9 empty lake beds, and dry channels were discovered on Titan’s surface, along
10 with vast expanses of equatorial dunes (Tomasko et al., 2005; Lorenz et al.,
11 2006b; Stofan et al., 2007; Hayes et al., 2008; Radebaugh et al., 2008; Hayes,
12 2016; Poggiali et al., 2016) that together point to a complex, dichotomous,
13 latitudinally-dependent climate (Mitchell and Lora, 2016). Yet despite these
14 advances, the nature of the connection between (sub)surface and atmosphere
15 remains elusive (Mitchell and Lora, 2016; Turtle et al., 2018b).

16 Huygens also provided the first *in situ* measurements of Titan, yielding
17 estimates of winds and measurements of temperatures from inside the atmo-
18 sphere at unprecedented precision, along with determination of the compo-
19 sition of the lower atmosphere (Bird et al., 2005; Fulchignoni et al., 2005;
20 Niemann et al., 2005; Niemann et al., 2010; Karkoschka, 2016). But data
21 from Huygens covered only one location during a single entry, and we still
22 have no direct measurement of the surface composition. Similarly, though
23 the Huygens temperature profile provided an invaluable confirmation of re-
24 trieved temperatures from other methods (Lindal et al., 1983; Schinder et al.,
25 2011), information on the variability of the atmosphere on diurnal and syn-
26 optic time scales is still lacking. These gaps prevent a full comprehension of
27 Titan’s meteorology, though modeling has helped to constrain our ignorance.

28 The increasing quality and complexity of data from Titan has motivated
29 the development and application of numerous general circulation models
30 (GCMs), with which many aspects of Titan’s climate system have been inter-
31 preted or inferred. Examples include the atmospheric superrotation, the cir-
32 culation and chemistry of the middle atmosphere, the tropospheric methane

33 cycle, the interactions of surface and atmosphere, as well as Titan’s paleo-
34 climate. But these studies have largely been carried out with independent
35 models (which also range considerably in complexity), and little attempt has
36 been made to understand or account for potential model biases or structural
37 differences, so many results may yet prove to be model-dependent.

38 Therefore, the purpose of this work is twofold: To synthesize and compare
39 detailed climate modeling results for Titan’s low-latitude climate, bench-
40 marked where possible by the *in situ* data, and at the same time to provide
41 context and expectations, based on multi-model results, for the Titan envi-
42 ronment at the time of arrival of the proposed Dragonfly rotorcraft lander.
43 The Dragonfly mission concept (Lorenz et al., 2018; Turtle et al., 2018a)
44 would represent a dramatic follow-up *in situ* investigation of Titan to Huy-
45 gens, enabling, among other things, atmospheric measurements and envi-
46 ronmental monitoring to improve our understanding of meteorology and the
47 methane cycle, and provide heretofore unparalleled constraints for climate
48 models. A thorough synthesis of the state of knowledge concerning Titan
49 climate models is thus a timely need.

50 Model intercomparison exercises have a long history. They have been
51 a major focus of the international Earth climate community over the past
52 two decades, most recently through the sixth phase of the Climate Model
53 Intercomparison Project (CMIP6; Eyring et al., 2016). In the context of
54 planetary climates, intercomparisons have also been carried out for Mars
55 and Venus (e.g., Lebonnois et al., 2013). But despite a relative wealth of
56 Titan GCMs, only very limited multi-model work or comparisons have been
57 attempted (Lorenz et al., 2012; Hayes et al., 2013; Griffith et al., 2014; Mc-

58 Donald et al., 2016), and benchmarking of simulations has been mixed. This
59 is particularly important for more complex models that attempt to simulate
60 Titan’s climate with high fidelity, as mechanistic understanding is more dif-
61 ficult in such cases and extended parameter explorations are impracticable.
62 Therefore, here we attempt a detailed intercomparison of Titan GCMs, and
63 obtain some high-level conclusions.

64 **2. Models and Methodology**

65 This paper presents an intercomparison of three general circulation mod-
66 els (GCMs) of Titan’s atmosphere, focusing in particular on the low latitudes
67 during the time following northern winter solstice, which is roughly equivalent
68 to the month of January on Earth. This region and time are chosen because
69 they correspond to the location and date of the descent of the Huygens probe
70 through Titan’s atmosphere in 2005, as well as to the arrival of the proposed
71 Dragonfly mission around 2034, almost exactly one Titan year later. We
72 focus specifically on the lower atmospheric and surface environment, and
73 compare the models’ climatological predictions for basic characteristics like
74 winds, temperatures, methane content, and, where possible, precipitation.

75 Three models are involved in this intercomparison, all of which are fully
76 three-dimensional and simulate the low to middle atmosphere of Titan; these
77 models have been widely used and are extensively documented in the litera-
78 ture (see Lebonnois et al., 2012; Lora et al., 2015; Tokano, 2019). Very brief
79 overviews of the models are given in the following subsection and in Table 1.
80 A fourth model, TitanWRF (Newman et al., 2016), was also invited but was
81 unable to participate.

Table 1: Titan GCMs used in this study

Model Name (Abbr.)	Resolution (lon×lat)	Topography	Methane Cycle	Moist Convection	Titan Years Run	Reference
IPSL Titan GCM (IPSL)	64×48	Yes	No ^a	No	10	Lebonnois et al. (2012)
Cologne Titan GCM (Köln)	32×24	Yes	Yes ^b	No	10	Tokano (2019)
Titan Atmospheric Model (TAM)	64×32	No	Yes ^b	Yes	20	Lora et al. (2015)

^aSurface methane mole fraction is forced to 0.0565 (Niemann et al., 2010).

^bSurface hydrology is treated with a simple bucket scheme.

82 The models used in this study represent some of the highest-fidelity simu-
83 lations of Titan’s climate, and are the most comprehensive Titan atmospheric
84 models available, though relative to Earth GCMs they can at most be consid-
85 ered models of intermediate complexity (as opposed to fully coupled climate
86 models or Earth system models). Other modeling frameworks have been
87 extensively used to study various aspects of Titan’s climate or atmosphere—
88 particularly using two-dimensional or axisymmetric domains (e.g., Mitchell
89 et al., 2006; Rannou et al., 2006; Cresspin et al., 2008) and/or simpler or more
90 idealized configurations (e.g., Mitchell et al., 2011; Schneider et al., 2012)—
91 but in this context we are most interested in models that attempt to simulate
92 Titan’s lower atmosphere with the highest level of realism.

93 *2.1. Model descriptions*

94 The Institut Pierre Simon Laplace Titan GCM (henceforth the IPSL
95 model) uses the finite-difference dynamical core of the Laboratoire de Météorologie
96 Dynamique (LMDZ5), with physical parameterizations for soil temperatures,
97 surface–atmosphere fluxes and turbulent diffusion, as well as Saturn’s gravi-
98 tational tide (see Lebonnois et al., 2012). The radiative transfer parameter-

99 ization, previously based on the model of McKay et al. (1989), has been up-
100 dated with a multiple-scattering code that employs correlated k -coefficients,
101 with a fixed vertical profile of haze opacity (Lavvas et al., 2010). A fixed
102 methane profile (Niemann et al., 2010) is used in the radiative transfer, and
103 the model does not fully simulate the methane cycle; instead, the surface-
104 level methane mole fraction is held constant everywhere, at a value of 0.0565
105 based on Niemann et al. (2010). The model uses the topography map of
106 Lorenz et al. (2013).

107 The University of Cologne (Köln) Titan GCM, henceforth the Köln model,
108 uses the grid-point dynamical core of the Aries/Geos GCM, updated to be
109 quasi-hydrostatic (Tokano, 2013). It parameterizes surface temperatures,
110 methane condensation, and surface–atmosphere fluxes with a moisture avail-
111 ability parameter of 0.5, as well as Saturn’s gravitational tide. A simple
112 bucket scheme is used for hydrology. For radiative transfer, the model uses
113 the scheme of McKay et al. (1989) with a multiplicative correction factor to
114 tune the heating/cooling rates to observed values (Tomasko et al., 2008c);
115 this guarantees accurate low-latitude temperatures, though the model greatly
116 underestimates latitudinal temperature gradients (Sec. 3.2; Tokano, 2019).
117 Topography based on the map of Lorenz et al. (2013) is included.

118 The Titan Atmospheric Model (TAM) uses the GFDL Flexible Modeling
119 System and employs the FMS spectral dynamical core plus component mod-
120 ules to parameterize unresolved physical processes. These include schemes
121 to compute moist processes, including convection, condensation, and precip-
122 itation of methane; surface and ground temperatures; surface–atmosphere
123 fluxes and turbulent diffusion; and full radiative transfer that uses a com-

124 bination of Cassini measurements (Tomasko et al., 2008a,b) and correlated
125 k -coefficients (see Lora et al., 2015). Saturn’s gravitational tide is not pa-
126 rameterized. The model employs a simple bucket scheme for hydrology, and
127 is run in a configuration with imposed surface methane at high latitudes
128 and liquid infiltration at low latitudes, which has been shown to reproduce
129 a number of observations of Titan’s hydroclimate (Lora and Mitchell, 2015;
130 Mitchell and Lora, 2016; Lora and Ádámkovics, 2017; Faulk et al., 2017).

131 *2.2. Methodology*

132 Multi-year simulations were run for all models (see Table 1) in order to
133 incorporate interannual variability. Due to the length of Titan years (29.5
134 Earth years) combined with the models’ high temporal resolution, record-
135 ing simulation results for every variable at every time step was impossible.
136 Therefore, we opted to save the average, maximum, and minimum of each
137 desired variable (at every model layer, where applicable) for each Titan day
138 (Tsol), allowing us to quantify the full range of diurnal to interannual vari-
139 ability but limiting the temporal resolution of the output. In addition to
140 the multi-year runs, a subset of additional short (10 Tsol) simulations were
141 run with sufficiently high output frequency to resolve diurnal variations (see
142 Sec. 5.3).

143 Each model was run in the configuration available and preferred by its
144 authors. Therefore, not all models fully simulate the methane cycle, and in
145 each case the surface methane reservoir is treated differently; only two mod-
146 els include the effects of topography, and only one model parameterizes moist
147 convection. These substantial differences imply that not all comparisons are
148 equal (or possible), though perhaps model agreement therefore suggests ro-

149 bustness. Nevertheless, this work represents the first dedicated intercompar-
150 ison of Titan GCMs, and we consider the cataloguing of major discrepancies
151 and areas of resemblance an important step toward characterizing the state
152 of the field.

153 **3. Global context**

154 Though our focus in this paper is the low-latitude environment, we begin
155 by comparing the general circulation and atmospheric structure of the lower
156 atmosphere as simulated by the three models, to provide a global context.
157 Figs. 1 and 2 show the zonal-mean zonal winds alongside meridional mass
158 streamfunctions and the zonal-mean temperatures for two contrasting times
159 of year, northern winter solstice ($L_S = 270^\circ$) and northern vernal equinox
160 ($L_S = 0^\circ$).

161 *3.1. Circulation*

162 The zonal winds simulated by the IPSL model increase from the surface
163 to lower atmospheric pressures at all latitudes, with particularly strong flows
164 concentrated around low latitudes that are a clear indication of substantial
165 superrotation (Fig. 1a,d). There is a local maximum in wind speed around
166 50 hPa and a region of lower speeds around 25 hPa that corresponds to the
167 low-latitude profile measured by the Huygens probe (Bird et al., 2005). There
168 is a slight latitudinal asymmetry in the winds at northern winter solstice
169 (Fig. 1a), which intensifies at the northern vernal equinox as peak wind
170 speeds shift northward (Fig. 1d).

171 The corresponding mean meridional circulation shows a near-global sol-
172 sticial Hadley cell at higher pressures, and a thermally indirect cell at the

173 pressures of the zonal wind local minimum. The circulation at equinox is
174 more complicated and shows considerable structure. In both hemispheres, a
175 direct circulation is apparent up to high latitudes, but punctuated by con-
176 fined regions of indirect flow. At lower pressures, the circulation is domi-
177 nantly clockwise, except at mid-northern latitudes where it reverses. Thus,
178 the meridional circulation in the lower atmosphere is separated from that of
179 upper levels. The latitudinal structure visible at higher pressures is due to
180 the presence of topography in this model; this is numerical noise that likely
181 does not accurately represent Titan, though its effects on momentum trans-
182 port and the superrotation are likely not significant (as seen in the zonal
183 winds).

184 The Köln model’s zonal winds are much lower in magnitude throughout
185 the atmosphere for both seasons, and do not show superrotation (Fig. 1b,e).
186 This model’s inability to reproduce Titan’s superrotating zonal circulation
187 has been noted since its inception (Tokano et al., 1999), and represents an
188 important deficiency. At higher latitudes (around 60° in the northern hemi-
189 sphere), wind speeds are slightly higher, but still do not reach 10 m s^{-1} .

190 On the other hand, the meridional circulation simulated by the Köln
191 model is much smoother than in the IPSL model, despite its also including
192 topography. At the solstice, there is a clear, global Hadley cell extending from
193 pole to pole and throughout the lower atmosphere (Fig. 1b), with a small
194 region of counterclockwise flow near the surface at mid-southern latitudes.
195 The values of the mass streamfunction are similar to the IPSL model’s around
196 100 hPa. At the equinox, the circulation shows two highly symmetric direct
197 cells, indicating rising motion precisely at the equator, with poleward flow

198 aloft (Fig. 1e).

199 The zonal winds simulated by TAM increase strongly with height through
200 the lower atmosphere, and reach values larger than 50 m s^{-1} at pressures
201 above 10 hPa at all latitudes equatorward of 60° , clearly showing robust
202 superrotation (Fig. 1c,f). Strong westerly winds extend closer to the surface
203 at higher latitudes. The solstitial winds are largely latitudinally symmetric
204 (Fig. 1c), and, as in the IPSL model simulation, peak winds shift northward
205 at equinox (Fig. 1f).

206 The meridional circulation simulated by TAM is very similar to that
207 from the Köln model, with some small differences. At solstice, a global
208 Hadley cell extends from south pole to high northern latitudes, but there is
209 a near-surface region of indirect circulation over the northern pole (Fig. 1f).
210 Similarly, roughly latitudinally symmetric direct cells occur at equinox (with
211 similar magnitudes as in the Köln model), but with near-surface indirect cells
212 over both poles. These shallow regions are the meridional manifestation of
213 baroclinic eddies at the high latitudes (Lora and Mitchell, 2015).

214 *3.2. Temperature*

215 Zonal-mean temperatures at northern winter solstice from the IPSL model
216 simulations are horizontally homogeneous at atmospheric pressures above
217 100 hPa, but have some latitudinal structure at lower pressures (Fig. 2a).
218 Temperature minima, centered around 100 hPa, are symmetric about the
219 equator (and coldest over the poles), but above that the highest temperatures
220 occur at low latitudes, while the northern polar atmosphere between 10–
221 30 hPa is much colder than its southern counterpart. At the northern vernal
222 equinox, the temperature distribution is quite similar, except that both the

223 100 hPa southern polar region and the 10–30 hPa northern polar region have
224 warmed by a few K (Fig. 2d). Thus, seasonal variations, though muted, are
225 perceptible.

226 In the case of the Köln model, there is essentially no temperature variation
227 with latitude at either season, meaning that temperature gradients are always
228 minimal (Fig. 2b,e). This results directly from the tuning of the model to
229 reproduce low-latitude observations. This approach has the side effect of
230 strongly damping temperatures deviations elsewhere, and thus suppresses
231 realistic temperature variations at other latitudes (and seasons).

232 The zonal-mean temperatures simulated by TAM show the largest (though
233 still small) latitudinal variations near the surface, and similar solstitial varia-
234 tions as in the IPSL model above (Fig. 2c). In particular, equatorial temper-
235 atures are warmest, and the northern polar region sees the coldest tempera-
236 tures. The differences are that the temperature minimum occurs at slightly
237 higher pressures (around 130 hPa), and vertical temperature gradients above
238 this level are somewhat stronger. The evolution of temperature into equinox
239 is also quite similar, showing warming of both polar regions and little change
240 at pressures above 300 hPa (Fig. 2f).

241 4. Low-latitude climatology and comparison to observations

242 We now focus specifically on the simulated climatology of the period
243 following northern winter solstice, approximately corresponding to $L_S = 290$ –
244 320° , as this is the only time of year on Titan for which *in situ* data exist. This
245 also permits predictions from the models for the same season in the future,
246 which would coincide with the arrival of the proposed Dragonfly rotorcraft

247 lander at Titan in 2034. In this subsection, the models' wind, temperature,
248 and methane mole fraction fields are compared to those *in-situ* data from the
249 Huygens probe, as a means of evaluating their fidelity, as well as in order to
250 elucidate the inter-model spread.

251 4.1. Winds

252 Zonal winds simulated by the IPSL model are shown in Fig. 3a. Near the
253 surface, these closely match the observed winds (Bird et al., 2005), with east-
254 erly winds of increasing magnitude with altitude below approximately 5 km.
255 Above this level, the simulated winds become increasingly more westerly, as
256 in the observed profile, but with insufficient shear; around 40 km altitude, the
257 model winds are roughly 10 m s^{-1} slower than the observations. Neverthe-
258 less, these simulations qualitatively reproduce the observed wind structure.
259 Meridional winds (Fig. 3b) are also in agreement with those inferred from the
260 movement of the Huygens probe (Karkoschka, 2016), including near-surface
261 northerlies and weak southerlies above that, with magnitudes close to zero
262 aloft. The exception is the weak northward winds suggested at 15 km that
263 exceed even the maximum winds at that level in the model.

264 The Köln model wind profiles are farthest from observations. The mean
265 value of the zonal wind stays near zero at all altitudes in contrast to the
266 substantial observed increase with altitude (Fig. 3e). The model's inabil-
267 ity to reproduce Titan's zonal winds is apparent here, with the mismatch
268 between simulations and measurements increasing with altitude. Near the
269 surface, weak easterly zonal winds closely match observations. Meridional
270 winds likewise match the inferred wind profile near the surface, though the
271 simulations seem to produce several vertical layers of winds in opposing direc-

272 tions, rather than just the two observed and simulated by the other models.
273 Above 5 km, the mean winds become increasingly northerly, instead of re-
274 maining at zero magnitude; in this case, however, the observed profile lies
275 within the 1st–99th percentile range (see section 5.1) of the model (except
276 around 15 km).

277 TAM wind fields agree best with the data (Fig. 3e, f), with the zonal wind
278 profile proving an excellent match. There is a slight ($1\text{--}3\text{ m s}^{-1}$) overestima-
279 tion of the zonal wind magnitude at most levels, but the wind direction and
280 magnitudes closely reproduce the observed profile, including the higher ver-
281 tical wind shear above 24 km than below. Additionally, the significant zonal
282 wind increase with altitude accurately captures Titan’s substantial equato-
283 rial superrotation. The model’s meridional winds also generally match the
284 observations albeit with weaker average magnitudes: in the boundary layer
285 there is mean southerly flow underlain by mean northerly flow, as inferred
286 from the movement of the Huygens probe. The average meridional wind
287 profiles simulated by TAM and IPSL are practically identical, though the
288 variability of the wind speed is considerably larger in the former. In this
289 case, the nonzero northward winds inferred at 15 km fall well within the
290 1st–99th percentile envelope of the simulation.

291 *4.2. Temperature and methane content*

292 All three models simulate equatorial temperature profiles (Fig. 3c,g,k)
293 that are in close agreement with that observed with the Huygens probe
294 (Fulchignoni et al., 2005). The IPSL model simulates temperatures that
295 are $\sim 1\text{--}2\text{ K}$ too cold between approximately 5 and 25 km altitudes, while
296 TAM produces temperatures that are slightly ($<1\text{ K}$) too warm around 10 km

297 altitude. The temperature structure from the Köln model is nearly identical
298 to the observed profile below 30 km altitude, but it should be noted that this
299 results from the fact that that model’s radiative heating rates are tuned to
300 reproduce the observations.

301 In all three models, temperatures near the temperature minimum (around
302 42 km) are slightly too warm. The same is true above these levels (not
303 shown). Bézard et al. (2018) suggest that a similar discrepancy in results with
304 a discrete ordinates one-dimensional model can be explained by adiabatic
305 cooling due to rising motion; however, it is not clear whether the GCMs
306 are all underestimating such a dynamical cooling, or whether these issues
307 instead result from systematic inaccuracies or uncertainties in the radiative
308 properties and/or profiles of atmospheric gases and haze.

309 The methane mole fraction profiles simulated by the models also quali-
310 tatively reproduce the measured profile (Niemann et al., 2010), though with
311 several discrepancies. In all cases, the lowermost ~ 5 km support constant
312 mole fractions of methane, which indicate a well-mixed, subsaturated region,
313 in agreement with the observation. Above this level, the mole fractions drop
314 sharply and then asymptote to a minimum level around 40 km. This is strong
315 evidence for methane being close to or at saturation between 5 and 40 km,
316 though the compositions of liquid or solid droplets, and therefore the exact
317 saturation values, are uncertain.

318 The IPSL model’s profile has mole fractions that are generally slightly
319 low (Fig. 3d), except at the surface. This is a result of the imposed rather
320 than prognostic simulation of near-surface methane content, in combination
321 with the model’s colder-than-observed temperature profile. The Köln model

322 and TAM both reproduce the observation remarkably well (considering, in
323 particular, their simple treatment of saturation values), though in both cases
324 the simulations slightly underestimate the observed methane mole fraction
325 below around 10 km altitude. Near the surface, the observed profile lies
326 within the 1st–99th percentile ranges of both models. In the Köln model,
327 the methane content agrees very well with the observations above around
328 40 km altitude, despite the slightly warm temperatures. In TAM, the too-
329 warm upper-level temperatures translate to methane mole fractions that are
330 somewhat too large above roughly 30 km.

331 *4.3. Latitudinal variations*

332 Figure 4 shows zonal-mean latitudinal profiles of 10 m altitude wind fields
333 (scaled from the models’ lowest levels), as well as surface temperatures, for
334 the early winter averaged over all simulated years for each model, over the low
335 latitudes. The profiles show some structure and there is general qualitative
336 agreement between models; there are also notable differences, but the range
337 of variability in each model is larger than the differences of the displayed
338 average profiles between models (see Sec. 5.2 for analysis of the variability).

339 The near-surface zonal winds (Fig. 4a) tend to shift from slightly westerly
340 to slightly easterly between southern and northern low latitudes. The IPSL
341 and Köln models indicate that the largest-magnitude average winds in the low
342 latitudes occur between 20 and 30°S, while at that location TAM simulates
343 near-zero average winds. Interestingly, the models coincide around 10°S, the
344 latitude of the Huygens landing. They also agree in the northern hemisphere,
345 where peak average easterlies occur in all cases between 10 and 25°N.

346 The simulated near-surface meridional winds (Fig. 4b) are northerly in

347 all models, in agreement with the expectation of cross-equatorial return flow
348 in the large-scale Hadley cell, with ascent in the southern hemisphere in this
349 season. In all cases, the peak average winds occur close to the equator, but
350 there is less agreement between the models regarding in which hemisphere.
351 In addition, the Köln model and TAM suggest that mean southerlies exist
352 poleward of around 25°S. In both of those cases, low-level convergence occurs
353 throughout the southern low-latitudes, while in the IPSL model this occurs
354 farther poleward.

355 Lastly, average surface temperature profiles (Fig. 4c) also coincide be-
356 tween models in predicting seasonal warmth in the low southern latitudes
357 relative to the northern hemisphere. The IPSL model and TAM agree re-
358 markably well in this range, though the maxima occur at different latitudes;
359 this is likely a consequence of the differences in the treatment of surface
360 methane and evaporation (see Sec. 6). The largest discrepancy is in the
361 profile from the Köln model north of 15°N, which drops substantially and
362 departs from the smoother profiles of the other two models; this behavior is
363 likely also the result of localized liquid buildup that is evaporating (Tokano,
364 2019).

365 **5. Variability**

366 *5.1. Variability in the atmospheric profiles*

367 In addition to average values, Figure 3 displays the range of 1st–99th
368 percentile, as well as absolute maximum and minimum values, in each simu-
369 lation, calculated over all years for $L_S \approx 290\text{--}320^\circ$. These statistics therefore
370 give a sense of the range of diurnal to interannual variability for the season

371 in question, as well as variability over longitude for IPSL and TAM (Köln
372 results are averaged in longitude). While the temporal and spatial scales
373 of variability are not individually resolved in this analysis, these statistics
374 represent the full range of values simulated by the models.

375 Predictably, wind profiles are considerably more variable than temper-
376 ature or methane mole fraction. This is because the thermal structure in
377 Titan’s lower atmosphere is largely controlled by radiative balance (McKay
378 et al., 1989), and radiative timescales there are long (Bézard et al., 2018); at
379 the same time, the methane mole fraction is strongly constrained by satura-
380 tion, and therefore by the invariable temperature.

381 For zonal winds, both the IPSL model and TAM simulate 1st–99th per-
382 centile ranges of approximately $1\text{--}3\text{ m s}^{-1}$ around the mean values, in the
383 former case increasing with altitude. The full range of simulated zonal winds
384 is only slightly larger, though in TAM the range near the surface increases to
385 between -5 and 5 m s^{-1} . This is likely a consequence of moist convection,
386 which intermittently alters the boundary layer and affects the wind struc-
387 ture. For the Köln model, the range of variability is similar to IPSL near
388 the surface, but increases with altitude and becomes increasingly asymmet-
389 ric about the mean. Around 40 km altitude, the range of simulated winds
390 is between around -0.5 and 5 m s^{-1} , while the mean value is approximately
391 1 m s^{-1} .

392 The relative variability of meridional winds shows more inter-model spread.
393 IPSL simulates meridional winds that, above 5 km altitude, never exceed
394 0.5 m s^{-1} in magnitude. Closer to the surface, the range increases, with
395 extrema reaching magnitudes of about 1 m s^{-1} . The spread of meridional

396 winds in the Köln model is intermediate between the models, ranging from
397 about $\pm 0.5 \text{ m s}^{-1}$ around the mean at roughly 5 km to $\pm 1 \text{ m s}^{-1}$ around
398 the mean at 40 km and near the surface. The increase of the variability
399 ranges with altitude above 5 km is also larger than for the other models,
400 as with the case for zonal winds. Finally, TAM simulates meridional winds
401 whose 1st–99th percentile range is approximately 1 m s^{-1} above and below
402 the mean everywhere, with minima and maxima that exceed this by about
403 another 0.5 m s^{-1} . Near the surface, these respective ranges roughly double.

404 The temperature variability in all cases is essentially negligible. This is in
405 agreement with the fact that all observed low-latitude temperature profiles of
406 Titan agree with each other remarkably well (Lindal et al., 1983; Fulchignoni
407 et al., 2005; Schinder et al., 2011). In the Köln model and TAM, the range
408 of temperatures around 40 km altitude is roughly 1 K around the mean
409 (Fig. 3c,g). And, in TAM, extreme values deviate from the mean by up to
410 several K near the surface, again likely as a result of relatively rare moist
411 convective events, which the other models do not simulate.

412 Lastly, modest variability of the methane mole fraction appears in all of
413 the models, with a few differences. As a result of its prescription at the
414 surface, the variability in methane in the IPSL model is negligible at low
415 altitudes, but the range increases with altitude to around 10 km, before
416 decreasing again. Similar subtle mid-level ranges are produced in the other
417 models, but in those cases the mole fraction below 5 km is most variable.
418 In TAM, the 1st–99th percentile range is still fairly close to the mean, with
419 more noticeable extrema, while in the Köln model the 1st–99th percentile
420 range is relatively large from the surface to ~ 5 km.

421 *5.2. Surface and near-surface variability*

422 Figure 5 displays the relative frequency distributions of surface and near-
423 surface variables closest to 10°S , the latitude of the Huygens probe’s landing,
424 simulated by each model (for $L_S \approx 290\text{--}320^{\circ}$). Each panel includes his-
425 tograms of minimum, maximum, and daily-mean (averaged over one Tsol)
426 values for each variable, illustrating the full range of variability for the sea-
427 son. The true frequency distribution of instantaneous values can be intuited
428 from these histograms, but is not available given our data volume limitations.

429 Simulated wind speeds from the lowest model layers are scaled to 10 m
430 winds for comparison. In all cases, these have magnitudes that rarely exceed
431 0.5 m s^{-1} , in agreement with observational estimates, though the distribu-
432 tions from the various models are quite different. Zonal winds average ap-
433 proximately zero, and both easterly and westerly winds occur in all models.
434 Meridional winds average negative (that is, northerly) values, which agree
435 with the inferred meridional wind from movement of the Huygens probe
436 (Karkoschka, 2016).

437 The near-surface wind distributions from the IPSL model are weakly bi-
438 modal, with a secondary frequency maximum close to 0.5 m s^{-1} in zonal
439 winds and close to zero in meridional winds (Fig. 5a,b). This is due to the
440 influence of topography, which induces wind variations in longitude. The dis-
441 tributions of mean, maximum, and minimum winds are closely overlapping,
442 so the range of diurnal-mean winds is representative of the full range. In
443 the Köln model, the wind frequency distributions are narrower than in the
444 other models (likely as a result of being zonal averages), and largely symmet-
445 ric, with a somewhat larger relative spread between minimum and maximum

446 than in the IPSL model (Fig. 5e,f). Lastly, the frequency distributions in
447 TAM are more spread out than the other models, with extrema that extend
448 relatively farther (Fig. 5i,j). These distributions are also not bimodal, since
449 the model does not include topography. And the distributions of meridional
450 winds are clearly positively skewed, though the highest frequencies of mean,
451 minimum, and maximum are all still negative (northerly). In all models, the
452 near-surface wind distributions agree very well with observational estimates.

453 Surface temperatures in the three models average between 93.5 and 94.0 K
454 (Fig. 5c,g,k), in agreement with the value measured by Huygens (Fulchignoni
455 et al., 2005). In the IPSL model, the frequency distributions have a secondary
456 maximum just below 93.5 K, though the total range from minimum to max-
457 imum is limited to 93–94.5 K (Fig. 5c). The distributions from TAM are
458 normal, and the range is similar with temperatures a few tenths of a degree
459 higher (Fig. 5k). In the Köln model, the distribution of mean surface tem-
460 peratures agrees with that of the other models, but does not overlap with the
461 distributions of minimum and maximum temperatures (Fig. 5g). The min-
462 imum temperatures are around 93.0 K, while the maximum temperatures,
463 which have a wider distribution, extend between 94.5–95.5 K, warmer than
464 the other models. The reason for this model’s considerably larger surface
465 temperature variability is not obvious, though it is potentially related to the
466 combination of topography and a low value of thermal inertia (IPSL uses a
467 value of $1000 \text{ J m}^{-2} \text{ s}^{-1/2} \text{ K}^{-1}$; Köln and TAM use a value of approximately
468 $335 \text{ J m}^{-2} \text{ s}^{-1/2} \text{ K}^{-1}$).

469 Substantial inter-model disagreement appears in the frequency distribu-
470 tions of boundary layer methane mole fraction (Fig. 5d,h,l). As discussed

471 above, this iteration of the IPSL model prescribes the surface-level methane
472 content, so its lack of variability and agreement with the observed value (Nie-
473 mann et al., 2010) are expected and non-predictive. The other two models
474 also agree well with the *in situ* measurement, which in both cases is the most
475 frequent value. In both models the distributions of mean, minimum, and
476 maximum mole fractions also largely overlap. On the other hand, the distri-
477 butions are very different between models, with the Köln model simulating
478 a much broader overall distribution (spanning 0.05–0.070) and TAM simu-
479 lating a slightly positively skewed but very narrow distribution (spanning
480 0.055–0.060), with the exception of extremely infrequent maxima that also
481 extend to 0.070. Whether these differences are attributable to the inclusion
482 of a topography map in the former model, moist convection in the latter, or
483 a deeper structural reason, is not clear.

484 Figure 6 shows the same relative frequency distributions of surface and
485 near-surface fields as Fig. 5, but for latitudes close to 10°N, as a means
486 of illustrating regional and seasonal variability across the low latitudes. In
487 many regards, these distributions are similar to those at 10°S, but there
488 are notable differences that are mainly attributable to the contrast between
489 summer and winter hemispheres. Zonal winds in all models are more negative
490 (more easterly), with the most common daily averages all falling below zero.
491 Meridional winds remain generally negative, and in addition the westerly
492 tails of the distributions are more negative, with no positive values simulated
493 by the IPSL and Köln models; furthermore, the distributions in TAM are
494 considerably less positively skewed.

495 The 10°N distributions of temperature are, in all models, slightly colder

496 than in the south, again as a result of crossing from summer to winter hemi-
497 spheres. In the IPSL model and TAM, the differences are slight, of less than
498 0.5 K (Fig. 6c,k); in the Köln model, the differences are of about 1 K, with
499 the maximum value approximately 2 K colder (Fig. 6g). In addition, the
500 latter model’s surface temperature distributions overlap considerably more
501 than in the south, with the range from minimum to maximum spanning
502 approximately 92.5–94 K (as compared to 92.3–95.6 K in the south).

503 The distributions of methane mole fractions are least different across the
504 equator. In the Köln model, the distributions extend to slightly lower values,
505 with minima of around 0.048 (Fig. 6h); in TAM, the distribution is not skewed
506 and the maximum values drop to approximately 0.060 (Fig 6l). (Again, the
507 IPSL surface-level mole fractions are externally constrained, so no differences
508 are expected or occur.)

509 *5.3. Diurnal Variations*

510 As a final intercomparison of the models’ simulated variability, a subset of
511 additional short (10 Tsol) simulations were run with sufficiently high output
512 frequency to resolve diurnal variations. These were run for the time of year
513 corresponding to $L_S \approx 305^\circ$. Fig. 7 displays the resulting near-surface fields.
514 In the IPSL and Köln models, the parameterized gravitational tide from Sat-
515 urn induces a very regular diurnal variation in the surface pressure of around
516 0.05% and 0.1%, respectively (not shown; see Tokano and Neubauer, 2002);
517 this effect is absent in TAM since the gravitational tide is not parameterized.

518 Both zonal and meridional 10 m winds in the IPSL model show almost
519 no diurnal variations (Fig. 7a,d), though there is a hint of strengthening
520 of northerly meridional winds in the local afternoon. Diurnal variations in

521 near-surface winds exist in the Köln model, but in general their amplitude
522 is considerably lower than day-to-day variability. Nevertheless, zonal winds
523 appear to be slightly more westerly in the early morning and more easterly by
524 early afternoon, while meridional winds appear more northerly in the morn-
525 ing than in the afternoon (Fig. 7b,e). In TAM, diurnal variations in winds
526 are less perceptible, though there is a slight trend toward lower-magnitude
527 winds, particularly in the zonal direction, during the afternoon (Fig. 7c).
528 It should be noted that in all of these cases, 10 m winds are scaled from
529 higher-altitude model levels, so subtle diurnal variations that would result
530 from closer coupling to the surface may be missed.

531 Surface temperatures show the most appreciable diurnal variations (Fig. 7g,h,i).
532 In the IPSL model, surface temperature changes are muted—consistent with
533 the use of a relatively high surface thermal inertia—but nonetheless notice-
534 able, peaking in the early afternoon. In the Köln model, surface temperature
535 variability is particularly strong in the low-latitude southern hemisphere, in
536 which some regions show up to 2 K changes between sunrise and afternoon
537 highs. In the northern hemisphere, the diurnal contrasts are smaller due to
538 the lower insolation, and the average day-to-night contrast is of about 1.5 K,
539 with temperatures peaking around 15:00 local time. In TAM, a similar diur-
540 nal pattern appears but with much more muted magnitudes of roughly 0.3 K,
541 despite the use of a relatively low surface thermal inertia. In all cases, the
542 diurnal variations are discernible despite day-to-day and regional variability.

543 Lastly, the boundary layer methane mole fractions show no clear signature
544 of diurnal variation. In the IPSL model, this results from the prescription
545 of surface methane mole fraction. In the Köln model, individual grid points

546 see substantial changes on short timescales, but there is no obvious diurnal
547 pattern (Fig. 7d). In TAM, the methane mole fractions are nearly constant
548 over the 10 Tsols analyzed (Fig. 7h). It should be noted that no precipitation
549 occurred at these latitudes in the models during these short simulations.

550 **6. Precipitation**

551 The detection and monitoring of tropospheric clouds on Titan has been
552 possible for about 20 years (Griffith et al., 2000; Brown et al., 2002; Roe et al.,
553 2005; Porco et al., 2005; Schaller et al., 2006, 2009; Turtle et al., 2011a; Ro-
554 driguez et al., 2011; Roe, 2012; Turtle et al., 2018b) and the Cassini mission
555 provided compelling evidence of copious precipitation events following large
556 cloud outbursts (Turtle et al., 2011b). Cloud activity at mid- and high lati-
557 tudes was common in the south during southern summer and early fall, and
558 was observed to pick up in the north at the end of the Cassini mission, near
559 the beginning of northern summer (Turtle et al., 2018b). In the most gen-
560 eral terms, this seasonal evolution was predicted and expected (e.g., Mitchell
561 et al., 2006), but the scarcity of low-latitude activity around the northern
562 vernal equinox and dearth of clouds throughout northern spring was a puz-
563 zle in view of the idealized picture of a convergence and upwelling region
564 traveling between the summer poles. Instead, consensus is emerging that
565 restricted (polar) reservoirs of liquid methane—in the form of observed lakes
566 and seas, but also of moist regolith and/or near-surface ground methane—are
567 necessary to reproduce both the distribution and relatively sporadic occur-
568 rence of clouds (Mitchell and Lora, 2016; Lora and Ádámkóvics, 2017; Faulk
569 et al., 2017; Turtle et al., 2018b). Nevertheless, the details of Titan’s hydro-

570 logic cycle are not yet fully understood, and represent a major open question
571 post-Cassini.

572 While Cassini and ground-based instruments detected many polar and
573 mid-latitude clouds around 2005, the environment measured *in situ* by Huy-
574 gens at low latitudes was of a stable atmosphere with relatively low near-
575 surface humidity (Fulchignoni et al., 2005; Niemann et al., 2010). These
576 observations are in line with the expectation of relatively dry equatorial con-
577 ditions that result from the long-term divergence of surface methane from the
578 low-latitudes (Rannou et al., 2006; Mitchell, 2008). Thin stratiform layers
579 of condensation were suggested based on the Huygens data (Tokano et al.,
580 2006), but it is clear that moist convective events and significant precipita-
581 tion would require a substantial perturbation from the measured profiles in
582 order to initiate (Barth and Rafkin, 2007, 2010). Whether this is indica-
583 tion of seasonally dry conditions, or the result of a recent precipitation event
584 stabilizing the atmosphere, is still under some debate.

585 Figure 8 shows the models’ seasonal distributions of zonal-mean precip-
586 itation, averaged over the simulated years (see Table 1). The results from
587 the IPSL model should be treated with some caution since the model’s cur-
588 rent configuration does not simulate a full methane cycle and includes pre-
589 scribed surface methane mole fraction so that the atmospheric humidity is
590 not freely simulated. The other two models do fully simulate the methane
591 cycle, but with some substantial differences. First, TAM includes a parame-
592 terization of moist convection (Lora et al., 2015), while the Köln model does
593 not (Tokano, 2019). Second, liquid methane produced by the large-scale
594 condensation scheme is allowed to re-evaporate through the atmosphere as

595 it falls in TAM, but is immediately precipitated to the surface in the Köln
596 model. Perhaps most importantly, the surface boundary in both cases is a
597 simple bucket scheme, but with very different assumptions: In TAM, the
598 polar regions have effectively inexhaustible surface methane and a modest
599 infiltration rate is implemented at low latitudes (Lora and Mitchell, 2015;
600 Lora and Ádámkóvics, 2017; Faulk et al., 2017), and any surface methane is
601 fully available to the atmosphere. In contrast, the Köln model is initialized
602 with a dry surface, there is no infiltration, and evaporation rates are scaled
603 with an arbitrary “availability factor” of 0.5 (Tokano, 2019). In addition,
604 the Köln model includes the topography map of Lorenz et al. (2013). The
605 resulting precipitation distributions, while similar in the most general terms,
606 are quite different in detail [compare Fig. 1 of Faulk et al. (2017) to Fig. 13
607 of Tokano (2019)].

608 The precipitation distribution from the IPSL model simulation follows
609 seasonal patterns, with peak precipitation rates occurring during summer-
610 time over both poles, and mild precipitation occurring at low and mid-
611 latitudes primarily in spring and summer (Fig. 8a). But precipitation tends
612 to occur quasi-continuously in many regions, which is inconsistent with Cassini
613 observations (e.g., Turtle et al., 2018b). In addition, there appear to be lati-
614 tudes of preferred precipitation, likely as a result of the numerical noise also
615 observed in the meridional streamfunction (Fig. 1d) and associated with the
616 implementation of topography.

617 Precipitation in the other models is more sporadic (Fig. 8b,c). In both
618 cases the most precipitation also falls over the polar regions, but occurs
619 roughly from mid-spring to mid-summer (that is, earlier in the season than

620 in the IPSL model). This is probably due to differences in the thermal inertia
621 of the surface used by the models. At lower latitudes, more precipitation
622 falls in the northern than the southern hemisphere in both models, though in
623 the Köln model simulation northern mid-latitude precipitation falls generally
624 year-round, while southern mid-latitude precipitation is almost negligible
625 and equatorial precipitation mostly occurs around northern summer solstice,
626 contradicting observations of clouds. In TAM, the seasonality is more obvious
627 and precipitation generally occurs in summer mid-latitudes and equinoctial
628 low-latitudes. It is worth noting that—assuming precipitation is a good
629 proxy for clouds—this model’s overall precipitation distribution compares
630 favorably with the distribution of observed clouds (Mitchell and Lora, 2016;
631 Faulk et al., 2017; Turtle et al., 2018b), and also exhibits intensity statistics
632 that correlate with the latitude distribution of alluvial fans (Faulk et al.,
633 2017).

634 Figure 9 shows time series of precipitation over the low latitudes averaged
635 over the simulated years (see Table 1). In the IPSL model, precipitation is
636 generally higher in the southern low latitudes than the northern, and precipi-
637 tation rates in both hemispheres increase, from a minimum just after summer
638 solstice, into the subsequent season. This means that the season of interest,
639 shortly after southern summer solstice ($L_S = 290\text{--}320^\circ$), sees near-minimum
640 levels of precipitation, which in the northern low latitudes in particular are
641 almost zero. In the southern low latitudes, the indicated persistent all-year
642 precipitation is incompatible with the presence of sand dunes throughout
643 the region, which require that the sand be at least sometimes dry. A robust
644 interpretation of the dunes would require a better sense of the character of

645 rainfall and subsequent drying, which is not treated in this model.

646 Low-latitude precipitation time series from the Köln model are shown
647 in the second row of Fig. 9. Average precipitation rates are always low
648 ($<2.5 \text{ mm Tsol}^{-1}$), and precipitation is quite continuous, which would im-
649 ply more continuous cloud coverage than observed. The maximum in both
650 northern and southern low latitudes occurs around northern summer solstice,
651 and this is followed by very low precipitation rates through winter solstice,
652 and then a slight increase during northern winter. The relative symmetry
653 between hemispheres at these low latitudes appears to result from the effects
654 of topography in this model, since the overall distribution of precipitation is
655 more asymmetric in simulations without it (Tokano, 2019). The model pre-
656 dicta a high frequency of at least weak near-equatorial precipitation between
657 $L_S = 290$ and 320° ; however, only a handful of low-latitude cloud events
658 (~ 5) were observed by Cassini during this period (Turtle et al., 2018b), con-
659 tradicting the simulations.

660 Low-latitude precipitation in TAM (Fig. 9, third row) behaves quite dif-
661 ferently. The results are averages over 20 simulated years so precipitation
662 appears more common than any individual year actually is, but nevertheless
663 there is clear indication of intermittency. In addition, the (mean) precipita-
664 tion rates extend up to 8.0 mm Tsol^{-1} , indicating that some precipitation
665 events are intense (Faulk et al., 2017), in these regions particularly around
666 northern summer solstice ($L_S \sim 90^\circ$). Lastly, precipitation is almost entirely
667 absent during northern winter, and in particular does not occur at all during
668 $L_S = 290\text{--}320^\circ$. (Indeed, in the 50 years of simulation used in Faulk et al.
669 (2017), no precipitation ever occurred at these low latitudes in this range

670 of L_S .) Its prediction of extended droughts with highly episodic rainfall is
671 also consistent with the existence of sand dunes. Therefore, TAM appears to
672 capture many aspects of observed Titan meteorology and its manifestations
673 in the landscape. Lastly, the presence of methane moisture at the Huygens
674 landing site (Lorenz et al., 2006a; Karkoschka and Tomasko, 2009; Niemann
675 et al., 2010) is compatible with the last rainfall having occurred there some
676 time before, as suggested in Fig. 9; the landing site was a stream bed, and
677 since ethane was also present, evaporation of methane may have been sub-
678 stantially depressed, therefore allowing near-surface methane to persist long
679 after precipitation.

680 7. Discussion and Conclusions

681 We have presented a comparison of three GCMs of Titan’s atmosphere
682 to each other and to *in situ* data. This generation of models, at the end of
683 the Cassini mission’s exploration of the Saturn system, contains numerous
684 improvements and modifications that have increased the realism and fidelity
685 of the simulations. In many respects the models agree well with the obser-
686 vations (and each other), though there remain a number of discrepancies, as
687 well as various aspects that make the models somewhat incommensurable.

688 It is nevertheless clear that the *in situ* measurements taken by the Huy-
689 gens probe are generally representative of Titan’s low-latitude atmosphere,
690 as also borne out by comparison to other observations (Lindal et al., 1983;
691 Schinder et al., 2011), and the three GCMs studied here do a reasonable job
692 of reproducing the observed low-latitude environment. The possible excep-
693 tion to this is the simulation of zonal winds above 5 km altitudes, which only

694 one of the models (TAM) reproduces in a quantitatively satisfactory way.
695 Furthermore, it is also apparent that, despite differences in the simulated
696 variability, the average expected values for low-level winds, surface temper-
697 atures, and, to a lesser extent, boundary layer humidities during northern
698 winter are fairly well constrained. Even the magnitudes of plausible variabil-
699 ity in the models are qualitatively similar.

700 Other specific findings, as well as some predictions for the low-latitude
701 environment during northern winter, have resulted from these comparisons:

- 702 • Titan’s strong zonal winds and concomitant superrotation remain chal-
703 lenging for models to reproduce precisely, but two models clearly attain
704 superrotating winds and one accurately reproduces the Huygens wind
705 profile
- 706 • The variability of zonal winds is small in all models, with 1st–99th
707 percentile values generally less than 5 m s^{-1} from the climatological
708 means for the lower atmosphere
- 709 • 10 m altitude winds rarely exceed magnitudes of 0.5 m s^{-1}
- 710 • 10 m altitude meridional winds generally blow from the north, in agree-
711 ment with expectations from large-scale circulation arguments
- 712 • Meridional winds are weak in models and observations in the boundary
713 layer, and nearly zero above, with 1st–99th percentile values less than
714 $\sim 1 \text{ m s}^{-1}$ from the climatological mean (though inter-model variability
715 is large in this regard)

- 716 • Atmospheric temperatures are highly invariable, in agreement with var-
717 ious prior observations
- 718 • Surface temperature variations are small but noticeable (in agreement
719 with Cottini et al., 2012), and are also the clearest indication of diurnal
720 variation, which shows substantial inter-model spread

721 The largest differences between the models concern the methane cycle,
722 though we emphasize that this is also the area in which structural differences
723 make intercomparison least meaningful. As has been shown in a range of
724 contexts, the amount and distribution of surface methane, and its availabil-
725 ity to the atmosphere, strongly affects the atmospheric humidity, moisture
726 transport, and precipitation frequency and distribution (Mitchell et al., 2006;
727 Mitchell, 2008; Mitchell et al., 2009; Tokano, 2009; Schneider et al., 2012; Lora
728 et al., 2015; Lora and Mitchell, 2015; Mitchell and Lora, 2016; Newman et al.,
729 2016; Lora and Ádámkovics, 2017; Faulk et al., 2017; Turtle et al., 2018b),
730 yet this remains one of the least constrained variables in Titan’s hydrologic
731 cycle. But this presents a considerable opportunity for future data—both in
732 the form of continued remote monitoring of cloud activity and with future
733 *in situ* measurements by the proposed Dragonfly mission concept—to make
734 high-impact contributions to our understanding of Titan and its climate sys-
735 tem.

736 This study offers further suggestion that future measurements of Titan’s
737 lower atmosphere and surface environments, at low latitudes, will on aver-
738 age not be very different from measurements made by Voyager, Cassini, and
739 Huygens. But the ability to characterize the range of variability of various

740 fields, as well as its timescales, will provide invaluable input to further dif-
741 ferentiating between successful and unsuccessful model configurations and
742 assumptions, and thereby improve our ability to assess underlying physical
743 mechanisms. In addition, future data that provides improved information
744 on regional variations will significantly enhance our understanding of the
745 processes affecting the connection between Titan’s surface and atmosphere.
746 Refining our process understanding through improved data and more de-
747 tailed interpretation with models will also inform characterization efforts of
748 Titan’s broader global climate, including the relationship between high and
749 low latitudes, the global methane cycle, and seasonal changes therein.

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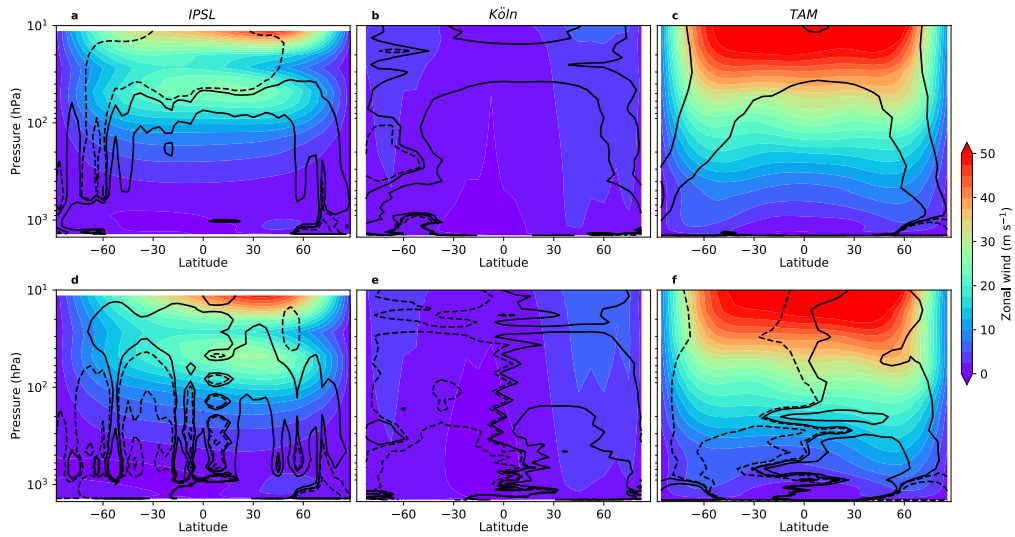


Figure 1: Zonal-mean zonal winds (colors) and meridional mass streamfunction (contours) for northern winter solstice (top row) and northern vernal equinox (bottom row). The columns show the results for different models. Positive zonal winds indicate westerlies (eastward). The contours show values of the streamfunction of ± 0.1 and $0.01 \times 10^9 \text{ kg s}^{-1}$; solid (positive) contours indicate clockwise motion and dashed (negative) contours indicate counterclockwise motion.

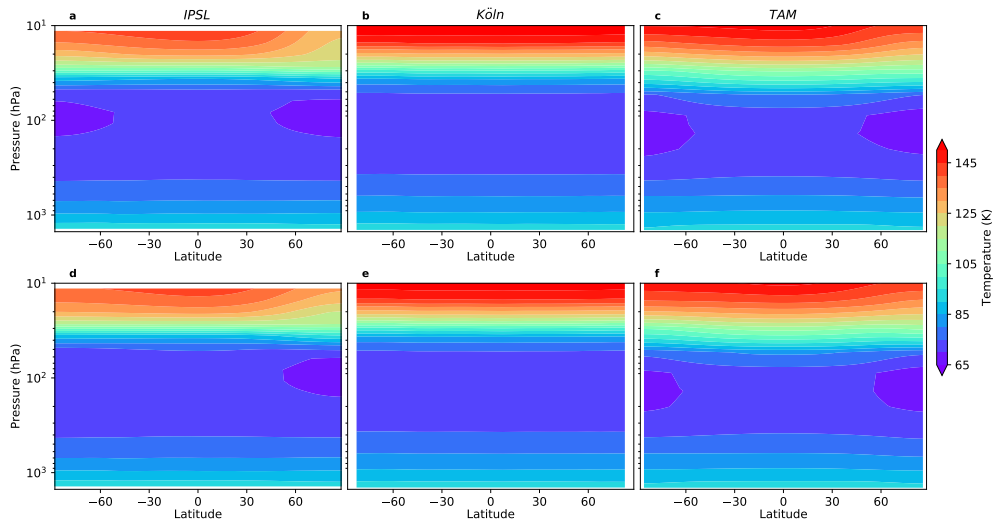


Figure 2: Zonal-mean air temperatures for northern winter solstice (top row) and northern vernal equinox (bottom row). The columns show the results for different models.

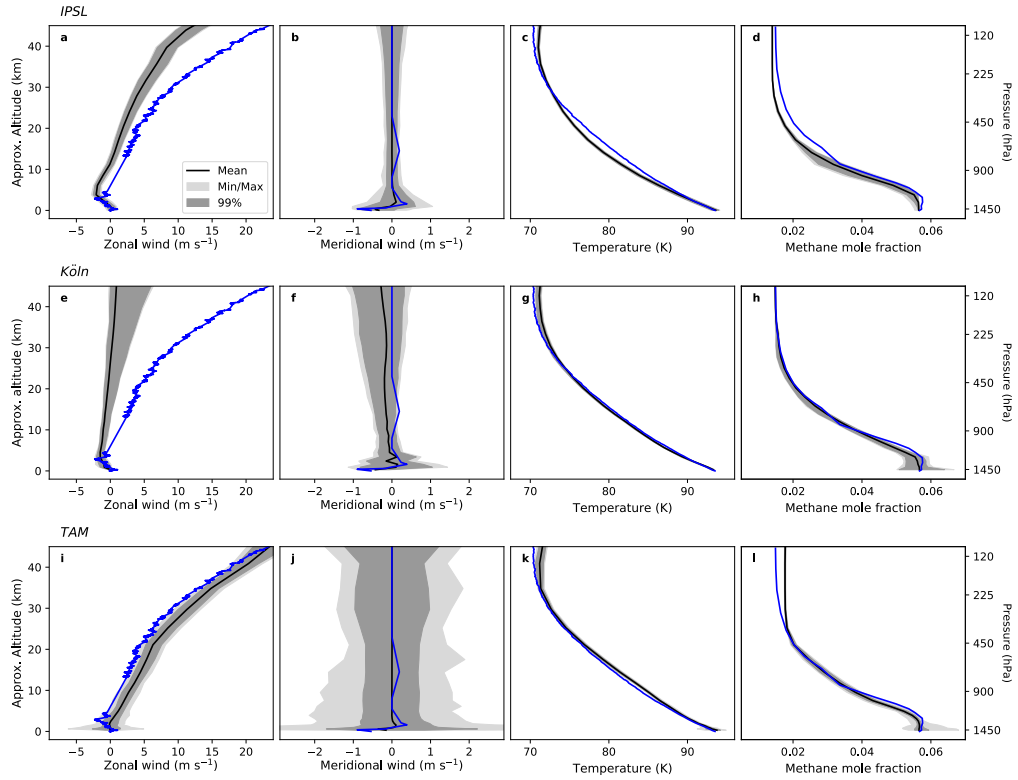


Figure 3: Vertical profiles of (columns from left to right) zonal winds, meridional winds, temperatures, and methane mole fractions simulated by the models and as measured *in situ* (blue curves) (Bird et al., 2005; Karkoschka, 2016; Fulchignoni et al., 2005; Niemann et al., 2010). Model results are shown at the grid point latitude closest to the Huygens probe landing at 10°S, considering all longitudes. Shaded regions show the ranges of minimum to maximum values (light gray) and 1st to 99th percentiles (dark gray), with the mean values for the season shown as black curves. The various rows show the results for different models. Positive zonal winds indicate westerlies (eastward), while positive meridional winds indicate southerlies (northward).

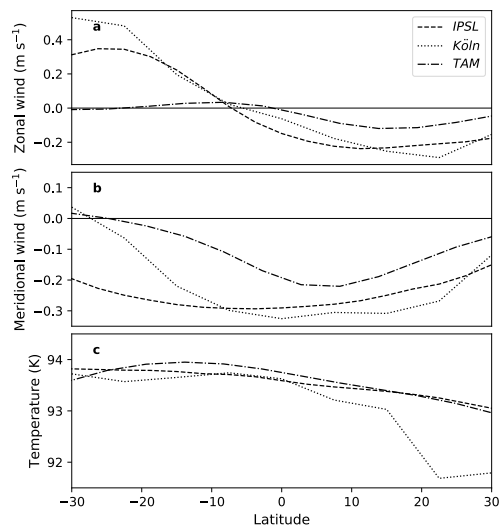


Figure 4: Latitudinal profiles of zonal-mean near-surface winds and surface temperatures, averaged between $L_S \approx 290\text{--}320^\circ$ and over all simulated years (see Table 1).

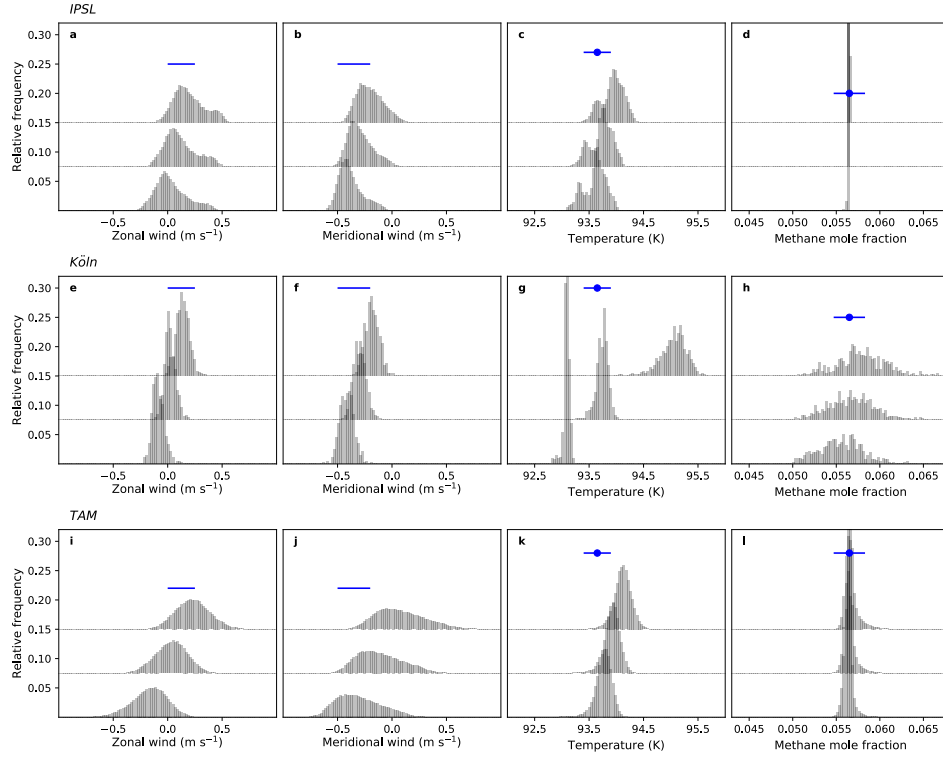


Figure 5: Histograms of (columns from left to right) near-surface winds, surface temperatures, and boundary layer methane mole fraction. Results are shown for model grid points closest to the Huygens probe landing at 10°S, considering all longitudes (results for the Köln model are zonal averages). Histograms for daily minimum, mean, and maximum values are shown in each panel—offset vertically from 0 to 0.15—to fully illustrate the range of variability in each simulation. For winds (left two columns), bottom-layer model results were scaled to 10 m above the surface. Methane mole fractions (right column) are shown for the bottom model layers, which are approximately at 40 m for the IPSL model, and 300 m for both the Köln model and TAM. In each panel, blue markers indicate observational estimates of winds (Lorenz, 2006; Karkoschka, 2007; Schröder et al., 2012) or measurements, with reported uncertainties, of temperature (Fulchignoni et al., 2005) and methane mole fraction (Niemann et al., 2010). The vertical position of these markers is arbitrary and chosen for clarity.

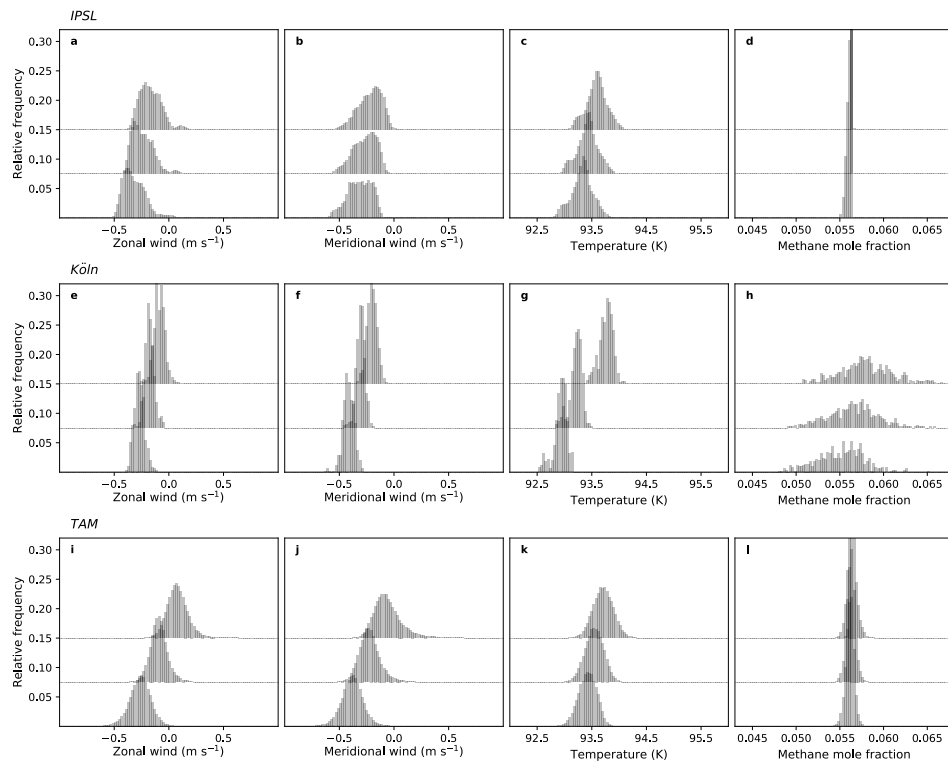


Figure 6: As in Fig. 5, but for the northern hemisphere (around 10°N).

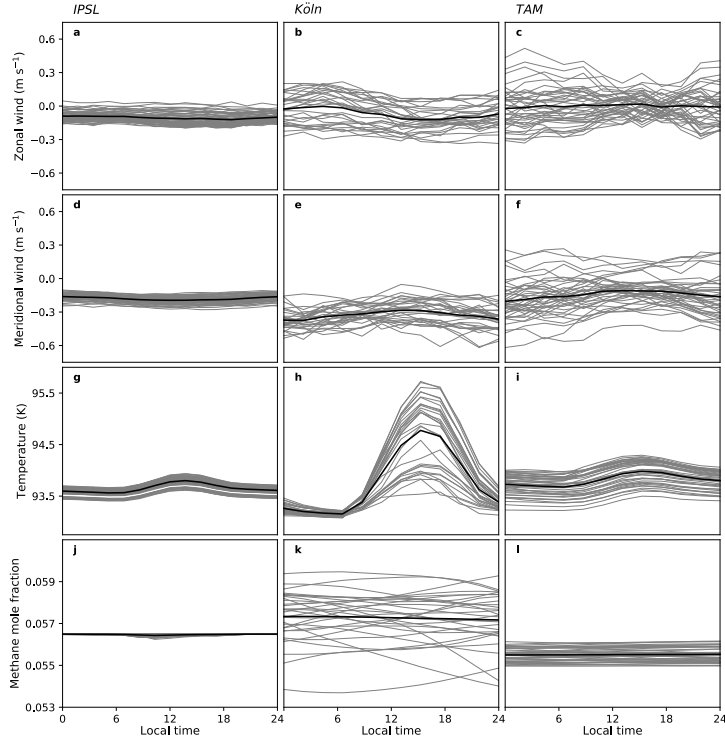


Figure 7: Diurnal variations (rows from top to bottom) of 10 m zonal wind, 10 m meridional wind, surface temperature, and boundary layer methane mole fraction at one longitude (the sub-Saturn point) over a 10 T_{sol} simulation around $L_S \approx 305^\circ$. Results are shown for all latitudes within 10° of the equator. Individual locations and days are shown in light gray curves, with the 10 T_{sol} and regional average shown in black. The columns show results for different models.

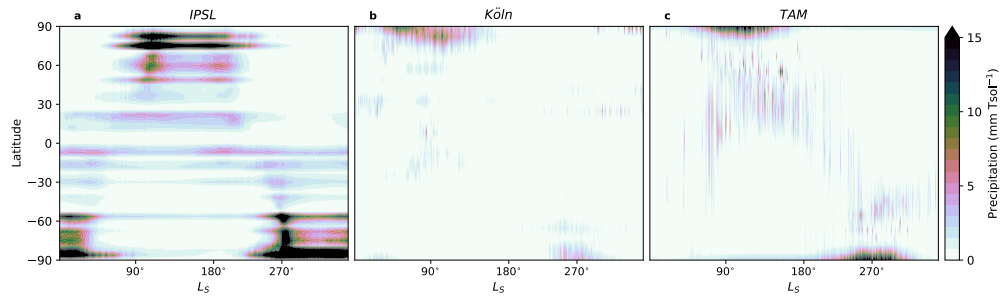


Figure 8: Zonal-mean climatological methane precipitation, starting from northern vernal equinox ($L_S = 0^\circ$), as simulated by the different models.

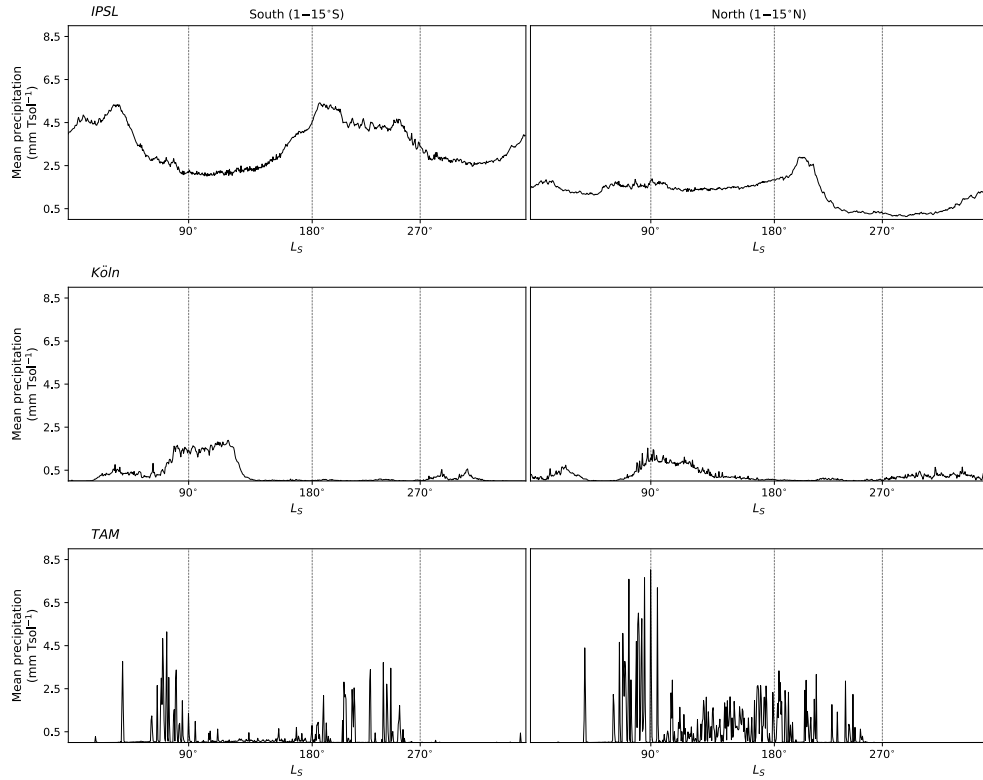


Figure 9: Time series of mean methane precipitation in the southern (left column) and northern (right column) low latitudes, starting from northern vernal equinox ($L_S = 0^\circ$). The various rows show the results for different models.