PLANETARY HYDROLOGY: IMPLICATIONS FOR THE PAST MARTIAN CLIMATE
AND PRESENT TITAN LAKE HYDROLOGY USING NUMERICAL MODELS OF
THE HYDROLOGIC CYCLES ON TITAN AND MARS

by

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ABSTRACT

The revelation, through the continued exploration of the Solar System, that Earth’s own hydrologic cycle is non-unique raises questions regarding the nature of hydrologic systems on planets and moons. Mars and Titan, the largest moon of Saturn, are prime candidates for exploring hydrology and hydrology-like systems in the Solar System. The past aqueous history of Mars tells a compelling story of perennial surface and subsurface hydrology following planet formation, and a drastic change in climate to the cold, dry planet at present-day. The nature of the past climate on Mars and climate change is debated, with some suggesting a cold, snow- and glacial-melt driven runoff and ponding cycle, while others favor a wetter past Mars consistent with liquid precipitation runoff and lake formation. Titan’s exotic atmospheric and surface chemistry, as well as its surface conditions allows liquid hydrocarbons to rain on the surface, flow as surface runoff through channels, infiltrate into the subsurface, form hydrocarbon lakes, and evaporate back in to the atmosphere, analogous to a hydrological cycle on Earth.

In this dissertation, I evaluate the past hydrology and climate at Gale Crater on Mars, the current site of the Curiosity rover and the present hydrocarbon-based hydrology of lakes on Titan using hydrological models of paleo-lakes. Modeling results are compared to the distribution of lakes on both Mars and Titan, as well as observed sedimentary layering and elevation of past lake stands in Gale Crater on Mars. On Titan, the distribution of methane rainfall, with a wet to dry transition from the poles to the equator, results in a precipitation-driven hydraulic head gradient and equatorward subsurface flow, inconsistent with the lower elevations observed at the poles. While no single model fits the observed lake distribution at the north polar region, a permeability that limits subsurface flow and the retardation of evaporation from the largest lake is required to match the observed lake distribution. On Mars, inferred lake stands and sediment
layering in Gale Crater provides constraints on the past climate of Mars and is largely consistent with a semi-arid climate. This work also shows that although sediment deposition will alter the crater topography, Gale craters unique location at the dichotomy boundary is advantageous for lake formation in a vertically integrated hydrologic system, even under dry climate conditions. In this thesis I show that using a hydrologic model that incorporates a subsurface, surface, and atmospheric component and observed lakes and paleo-lake environments on extra-terrestrial bodies, the past and present subsurface and atmospheric conditions on data sparse worlds can be constrained.
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CHAPTER 1

INTRODUCTION

As the sun sets on the Cassini mission’s survey of the Saturnian system, a look back on last 20 years of space and planetary exploration has seen a Mercury orbiter, a Venus orbiter, nine Lunar orbiters, seven Mars orbiters and four landers / rovers, two dwarf planet missions, numerous comet and asteroid missions, one Jupiter system orbiter, and Cassini’s mission to the Saturnian system and a lander on Saturn’s moon, Titan. One of the most profound discoveries has been the revelation of past and present active hydrology on Solar System bodies other than Earth, namely Mars and Titan (e.g., Masursky, 1973; Porco et al., 2003). Understanding the nature and fate of the hydrology and climate of these other bodies, and how these relate to Earth, has been a goal of scientists since the discovery of this extra-terrestrial hydrology.

Some of the earliest evidence of ancient water on Mars came from the Mariner 9 mission, which revealed the existence of ancient dry channels on Mars (Masursky, 1973) ranging from large outflow channels (Baker and Milton, 1974) to smaller dendritic valley networks (Carr and Clow, 1981). This provided evidence for a past climate much different than that at present day. The broad distribution of dendritic valley networks in the Noachian-highland terrain led to the idea that Noachian Mars was warmer and wetter (e.g., Craddock et al., 1997; Craddock and Howard, 2002). Later observations and improved image resolution allowed for more detailed analysis of these features, finding that most valley networks are more mature than previously thought and were likely formed in a climate that supported precipitation-driven runoff (Hynek and Phillips, 2003). Further evidence of a wetter early Mars came from spectral observations of hydrated minerals on the surface (e.g., Christensen et al., 2000; Bibring et al., 2006; Mustard et al., 2008; Murchie et al., 2009) associated with sedimentary rocks (Malin and Edgett, 2003).
Previous work aimed at determining the past climate focused on morphological characteristics (Howard, 2007) and discharge estimates (Hoke et al., 2011) of Martian channels. While hydrological modeling work has been done on Mars, linking the subsurface hydrology with the location of sediment deposits on the surface (Andrews-Hanna et al., 2007; Andrews-Hanna and Lewis, 2011), none of the aforementioned studies have linked the observed sediment deposits, lake stand indicators, and channels with the past climate in a vertically integrated Martian hydrological cycle.

Beyond Earth, Titan, the largest moon of Saturn, is the only body in the Solar System that currently contains stable liquid at the surface. Pre-Cassini Voyager 1 observations and theoretical studies determined that Titan could host liquid hydrocarbons, in particular methane, on the surface and possibly an active hydrocarbon-based hydrological cycle (Sagan and Dermott, 1982). Earth-based observations noted seasonal changes in the atmosphere that suggested a possible exchange with the surface (e.g., Lorenz et al., 1999). The arrival of Cassini at Titan, revealed lakes primarily concentrated at the north polar region (Stofan et al., 2007; Hayes et al., 2008) and fluvial dissection of the surface due to runoff of liquid hydrocarbons (e.g., Burr et al., 2006; Perron et al., 2006; Jaumann et al., 2008; Lorenz et al., 2008a; Burr et al., 2009; Black et al., 2012; Langhans et al., 2012). Furthermore, repeat observations noted changes in the surface radar brightness that appeared to be attributed to methane precipitation, and subsequent evaporation (Turtle et al., 2009; Turtle et al., 2011), indicate an interaction with the atmosphere, surface and possibly the subsurface. Thus far, only a handful of studies have investigated the role that the subsurface plays in the formation and stability of lakes on Titans surface (Hayes et al., 2008; 2012). Numerical modeling thus far has been limited by the sparse Titan topographic dataset and limited observations of temporal changes at the surface.
Despite the abundance of observational evidence for a past hydrological cycle on Mars and an active hydrological cycle on Titan, little has been done to investigate the role these hydrological cycles play in the past climate, the deposition of sediment, and the formation and distribution of lakes. In order to bridge the gap between observation and the hydrological system, this dissertation focuses on using numerical hydrological modeling of lakes on both Titan and Mars in order to link the surface, subsurface and atmosphere with surface and atmospheric observations. Specifically, this work focuses on hydrocarbon lakes at the north polar region of Titan (Chapter 2) and Gale Crater on Mars (Chapters 3 and 4), which holds evidence for a past aqueous environment. The primary hydrological model used in this work is developed in Chapter 2 and is then applied toward understanding the hydrology of Titan. Chapters 3 and 4 use the hydrological model, but with parameters more appropriate for Mars. Chapter 3 focuses on the lake hydrology and the climate during the early to late Hesperian after the formation of the sediment mound in Gale. Lake stands inferred from fan deposits in Gale are used to constrain the climate during this period. Chapter 4 examines the hydrologic conditions during the formative period of Gale crater, prior to the deposition of the central sediment mound. Lake deposits are used to constrain the climate during this period and crater infill models investigate the influence of sediment deposition on the formation of lakes in Gale. In using a numerical representation of these physical systems and the observed lakes and past aqueous environments, I show that when investigating the past climate and current hydrology of Mars and Titan respectively, it is important to consider the full hydrological cycle.
CHAPTER 2

THE INFLUENCE OF SUBSURFACE FLOW ON LAKE FORMATION AND NORTH POLAR LAKE DISTRIBUTION ON TITAN

Modified from a paper published in *Icarus* 2016

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**Abstract.** Observations of lakes, fluvial dissection of the surface, rapid variations in cloud cover, and lake shoreline changes indicate that Saturn’s moon Titan is hydrologically active, with a hydrocarbon-based hydrological cycle dominated by liquid methane. Here we use a numerical model to investigate the Titan hydrological cycle – including surface, subsurface, and atmospheric components - in order to investigate the underlying causes of the observed distribution and sizes of lakes in the north polar region. The hydrocarbon-based hydrological cycle is modeled using a numerical subsurface flow model and analytical runoff scheme, driven by a general circulation model with an active methane-cycle. This model is run on synthetically generated topography that matches the fractal character of the observed topography, without explicit representation of the effects of erosion and deposition. At the scale of individual basins, intermediate to high permeability \((10^{-8}-10^{-6} \text{ cm}^2)\) aquifers are required to reproduce the observed large stable lakes. However, at the scale of the entire north polar lake district, a high permeability aquifer results in the rapid flushing of methane through the aquifer from high polar latitudes to

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dry lower polar latitudes, where methane is removed by evaporation, preventing large lakes from forming. In contrast, an intermediate permeability aquifer slows the subsurface flow from high polar latitudes, allowing greater lake areas. The observed distribution of lakes is best matched by either a uniform intermediate permeability aquifer, or a combination of a high permeability cap at high latitudes surrounded by an intermediate permeability aquifer at lower latitudes, as could arise due to karstic processes at the north pole. The stability of Kraken Mare further requires reduction of the evaporation rate over the sea to 1% of the value predicted by the general circulation model, likely as a result of dissolved ethane, nitrogen, or organic solutes, and/or a climatic lake effect. These results reveal that subsurface flow through aquifers plays an important role in Titan’s hydrological cycle, and exerts a strong influence over the distribution, size, and volatile budgets of Titan’s lakes.

2.1 Introduction

The surface of Saturn’s largest moon Titan has been extensively modified by processes related to liquid hydrocarbons on the surface. Similar to water on Earth, methane at Titan’s surface is near its vapor point and exists in both the gas and liquid phases, suggesting that it is the primary constituent in the hydrocarbon-based hydrological cycle. Ponded liquid on Titan’s surface is primarily concentrated around the north polar region (Stofan et al., 2007) in lakes with variable morphology (Hayes et al., 2008). Lakes with steep-sided and smooth shorelines with no observable fluvial source may be indicative of karst or seepage morphology (Mitchell et al., 2008; Cornet et al. 2015) and can be either liquid filled or empty (Hayes et al., 2008). Larger lakes and seas with irregular shorelines, appear to be located in topographic lows (Stiles et al., 2009) at the terminus of fluvial features (Cartwright et al., 2011; Langhans et al., 2012). Surface modification by fluvial dissection (Burr et al., 2006; Perron et al., 2006; Jaumann et al., 2008;
Lorenz et al., 2008a; Burr et al., 2009; Black et al., 2012; Langhans et al., 2012) is driven by precipitation reaching Titan’s surface (Tokano et al., 2001).

The current orbital obliquity of the Saturn system of 26.7° causes seasonal variations in solar insolation (Lorenz et al., 1999; Stiles et al., 2009), which in turn drives seasonal variations in the polar precipitation and evaporation rates with an annual period of 29.5 years (note that the unit of time “year” will refer to an Earth-year throughout this work, unless specified as a Titan-year). During northern summer, high solar insolation results in atmospheric upwelling at the north polar region, causing rapid cooling of air parcels and condensation of methane to form clouds and precipitation. Observational evidence of cloud formation (Porco et al., 2005) and subsequent darkening of the surface in the south polar region during southern summer (Schaller et al., 2006; Turtle et al., 2011) suggests that convective storm systems bring precipitation to the surface. General circulation models (GCMs) predict increases in the evaporation and precipitation rates (Tokano et al., 2001; Mitchell et al., 2006; Schneider et al., 2012; Lora et al., 2014; Newman et al., 2015) due to high solar insolation at the polar regions during the summer and spring seasons.

Temporal changes, thought to be due to changes in the distribution of methane on the surface, have been observed at the south polar region and the tropics on Titan. Changes in the surface albedo in the south polar region and tropics are thought to be due to precipitation events and subsequent evaporation (Turtle et al., 2009; Turtle et al., 2011). Evidence for temporal changes in the distribution of lakes includes possible present-day changes in the location of the shoreline of Ontario Lacus in the south polar region (Turtle et al., 2011), and dry lakebed morphologies in the north polar region (Hayes et al., 2008) suggesting long-term changes in lake stability. The proposed shoreline change at Ontario Lacus suggests an average loss rate of ~1
m/yr for lakes in the south polar region based on an average shoreline recession of ~2 km (Hayes et al., 2011). However, alternative interpretations have suggested that the spatial resolution of the instrument used was inadequate for determining any measurable shoreline recession (Cornet et al., 2012). While dry lakebed morphologies (Hayes et al., 2008), inferred long-term lake level changes (Stofan et al., 2007; Lucas et al., 2014) indicate long-term changes in lakes, and transient lake features have been observed between subsequent Cassini flybys (Hofgartner et al., 2014), shoreline change in filled northern lakes during the Cassini mission has yet to be observed (Hayes et al. 2011). Based on predicted evaporation and precipitation rates, Mitri et al. (2007) placed theoretical upper limits on lake recession, predicting up to 30 km/yr for shoreline slopes of 0.1% based on altimetric profiles (Elachi et al., 2005). However, that early study required several simplifying assumptions, including constant slope, constant wind speed, and a lack of methane supplied from the surrounding watershed. Hayes et al. (2008) modeled temporal changes in lake area for lakes perched above an aquifer and separated from it by an unsaturated zone, specifically focusing on the timescales for lake disappearance and the influence of aquifer permeability on subsurface and atmospheric exchange of liquid methane. They found that for permeabilities less than $10^{-6}$ cm$^2$, lake recession becomes limited by the evaporation rate and occurs on the order of seasonal timescales. Higher permeabilities, on the order of $10^{-6}$ cm$^2$ to $10^{-5}$ cm$^2$, were found to be more consistent with seepage morphology lakes at the north polar region based on the size of these lakes.

Thus, previous work has revealed evidence for atmosphere-surface exchange of volatiles, transport of liquid over the surface, the existence of stable lakes and seas, and possible limited temporal changes in lakes and seas on Titan. Although there is clear evidence for an active hydrological cycle, significant work remains to be done to understand the nature of that
hydrological cycle. In particular, the role of subsurface flow in unconfined aquifers, the influence of this flow on the stability and distribution of lakes, and the properties of Titan aquifers are poorly understood. Subsurface hydrology on regional and global scales on Earth and Mars are important for understanding the distribution and activity of groundwater. A lack of ground truth observations of Titan’s subsurface hydrological properties necessitates comparison of hydrologic models with the observed distribution and behavior of lakes on Titan. While similar in many respects to the water-based hydrological cycles on Earth and Mars, Titan’s hydrocarbon-based hydrological cycle involves a fluid with a lower viscosity and density (depending on the assumed fluid composition), that flows under the influence of a weaker gravitational acceleration on a surface with lower relief, and driven by a longer seasonal cycle. These differences highlight the importance of theoretical studies of Titan’s hydrological cycle.

In this chapter, we model the full methane-based hydrological cycle of Titan, including atmospheric, surface, and subsurface components. In order to investigate lake behavior on both basin and polar scales, we combined a numerical subsurface hydrological model with an analytical surface runoff model, driven by the outputs from a general circulation model. We compare the results to observations of the distribution and sizes of lakes in order to constrain subsurface properties. In Chapter 2.2, the Titan hydrological model is described in detail. In Chapter 2.3 we use basin-scale hydrological models driven by precipitation and evaporation rates from a general circulation model at individual latitudes to investigate the behavior of hydrology on Titan at the scale of individual basins. The basin-scale model is then expanded to a polar model extending from the pole to mid-latitudes in Chapter 2.4, with latitudinally varying precipitation and evaporation rates. These polar models allow us to investigate the influence of different hydrological parameters on the formation of large seas and the distribution of lakes at
the north polar region. The results are compared with observational constraints in the form of the observed lake distribution at the north polar region, and implications for Titan hydrology are discussed in Chapter 2.5.

2.2 Methodology: Modeling Titan Hydrology

In order to investigate lake behavior on both basin and polar scales, we developed a numerical model that incorporates the atmospheric, surface, and subsurface components of Titan’s hydrological cycle. The model was run on two-dimensional grids representing the surface topography of either an individual basin or the entire north polar region (Chapter 2.2.1). The amount of methane that either recharges the aquifer or channelizes as surface runoff was determined from the outputs of a general circulation model (Chapter 2.2.2) using an Earth-based scaling relationship dependent on the precipitation and evaporation potential (defined as the evaporation rate that would occur from a standing body of liquid methane) (Chapter 2.2.3). The subsurface flow was modeled using a finite-difference approximation to the groundwater flow equation with parameters appropriate for Titan (Chapter 2.2.4). Surface runoff was modeled using a linear reservoir model with parameters appropriate for terrestrial basins (Chapter 2.2.5). The model allowed lakes to form and evolve naturally as a result of the balance between the surface, subsurface, and atmospheric fluxes of methane (Chapter 2.2.6).

2.2.1 Topography

Both the gravitationally driven flow of fluids in the subsurface and the surface runoff are dominated by the effects of the surface topography. For this study, we use topography derived from the overlapping antenna beams of the Cassini synthetic aperture radar (SAR) for all Titan flybys up through the T84 flyby (Stiles et al., 2009). The total relief (referenced to the geoid) on Titan is ~2.5 km, while the relief on smaller scales relevant to the basin-scale hydrological
modeling in this study is ~1.4 km (Figure 2.1). In an unconfined aquifer, where the methane
table generally follows a diffused representation of the surface topography, low relief will cause
lower hydraulic gradients, thereby decreasing the rate of subsurface flow. Furthermore, as a
result of the low relief on Titan, fluid at greater depths will not significantly affect the surface
hydrological cycle.

The limited surface topography available for Titan (Stiles et al., 2009; Figure 2.1) is
inadequate for direct hydrological modeling of specific regions, with the exception of a few areas
for which stereo-SAR coverage exists (Kirk et al., 2013). We circumvent this limitation using
synthetic topography. Surface topography is often approximately self-affine, behaving in a
fractal nature (Mandelbrot et al., 1982). This self-affine nature allows for a statistical measure of
surface topography to be calculated from individual topography swaths across specific regions of
interest. The fractal nature of the surface topography can be quantified by the Hurst exponent
(Campbell, 2002), which describes how the variance in elevation of points separated by a given
distance depends on that distance. The Hurst exponent is calculated as half the slope of the log of
the distance-dependent variance in elevation as a function of the log of the distance. The Hurst
exponent calculated from two-dimensional SAR topography (Stiles et al., 2009) or radar
altimeter swaths (Elachi et al., 2005) can be used to generate synthetic fractal terrain that is
statistically similar to the observed landscape, but is sampled over a dense two-dimensional grid
suitable for hydrological modeling. The variance in elevation as a function of the distance
between two points in a single SAR topography profile was determined and binned at different
distance scales. The slope (Hurst exponent) and intercept of the best-fit line to the log of the
variance in elevation as a function of the log of the distance were then used as inputs to the
fractal generation algorithm. In this representation, a slope of 1 indicates perfectly self-affine
behavior. While processes such as erosion, deposition, and dissolution of material can produce surfaces that are not self-affine, some large north polar lake shorelines appear fractal (Sharma and Byrne, 2010), indicating that the surrounding topography can be approximated by a fractal surface. Small steep-sided depressions that make up the majority of lakes at the north polar region appear to have rounded shorelines that formed due to dissolution. In this work, we focus on the large north polar seas and assume a fractal terrain surrounding these large basins. Our analysis below of individual topography swaths further supports the fractal nature of the topography at scales relevant to this work. Future work will consider departures from a fractal landscape as a result of the modification of the surface by erosion, deposition, and dissolution.

Figure 2.1. Global coverage of the synthetic aperture radar topography (Stiles et al., 2009) overlain on global Cassini Imaging Science Subsystem (ISS) mosaic (image credit: NASA/JPL-Caltech/SSI, http://astrogeology.usgs.gov/search/map/Titan/Cassini/Global-Mosaic/Titan_ISS_P19658_Mosaic_Global_4km). The red box surrounds Ligeia Mare, the region of interest for the basin-scale models, and the black box encompasses the north polar lake region of interest for the polar models.
We focus our basin-scale topographic analysis on an area in the north polar region in the vicinity of Ligeia Mare. The fractal analysis in this region yielded a two-slope fit for the Hurst exponent. At distances less than 10 km, a slope and intercept of 0.724 and -0.854 were calculated, while at distance greater than 10 km a slope and intercept of 0.272 and 0.944 were calculated (Figure 2.2). The greater slope of the fit at distances less than 10 km indicates that the topography displays a more self-similar nature than at distances greater than 10 km. We note that the effective footprint of the SAR topography of ~9 km (Stiles et al., 2009) may bias the Hurst exponent at scales smaller than this, but our model results are insensitive to the relief at these finer scales. The north polar region is relatively rougher than other regions on Titan (Stiles et al., 2009) possibly due to the enhanced hydrological activity and erosion at high northern latitudes. For the polar models, topographic statistics were determined at 15° latitude increments from 45° N to the pole. As in the case of the topography in the region surrounding Ligeia Mare, a two-slope fit with the break at 10 km provided the best fit to the data. Slope values for distances greater than 10 km suggest a decrease in self-similarity moving north towards the pole with slope values of 0.384, 0.114, and 0.113 for latitudes between 45° and 60°N, 60° and 75°N, and 75° and 90°N respectively. At distances less than 10 km, slope values remained consistently around 0.7, again suggesting a more self-similar terrain at shorter distances.

Synthetic topography was generated using a diamond-square fractal generation algorithm (Fournier et al., 1982). This method takes an initial square, defined at the four corners of the topographic domain, and generates a topographic value in the center of the square using a random perturbation from the mean of the four corners that depends on the distance from the computational node to the corners, and the slope and intercept derived from the distance-dependent variance based on the SAR topography profiles. The diamond step utilizes the newly
generated center elevation and the four corner pairs (NW-NE, NE-SE, SE-SW, SW-NW) to
generate topography at the central north, east, south, and west nodes, respectively. This process
is repeated until a two dimensional synthetic topographic map has been generated (Figure 2.3).

Figure 2.2. The log of the distance dependent deviation as a function of the log of the distance
from SAR topography over the region surrounding Ligeia Mare (solid circles) and the best fit
line to the data (dashed lines). A two-slope fit provides the best fit to the data with a break at 10
km.

For the basin-scale models, a purely synthetic fractal topography grid with a central
depression was used to model a closed topographic depression. Because our primary interest is in
the nature of the lake, a central depression was imposed on the grid rather than allowing it to
form at a random location. An initial 3x3 grid of points was determined by generating a coarse
grid landscape at a distance spacing equivalent to half the length of the model domain and then
selecting a central low point such that the distance dependent variance was preserved. For the
polar models, in order to provide a direct representation of the north polar region, an
interpolation algorithm using the SAR topography was included in the fractal generation (Figure
2.4a). This algorithm used an inverse distance-squared weighting interpolation to generate a 2×2
grid of points from the existing topography data that acted as the initial inputs to the fractal
algorithm. In the subsequent fractal generation algorithm, if the distance from a computational
node to an observed elevation point was less than the distance to the four square or diamond points, the data point was used in the fractal algorithm along with the corner points with an inverse distance-squared weighting. This method fits the SAR topography data while preserving the fractal nature of the landscape (Figure 2.4b).

The fractal algorithm alone, though, does not account for the location of the large basins containing seas at the north polar region. Thus, these basins were imposed in the polar topography models. Using the Cassini radar map of the north polar region, the four largest seas (Kraken Mare, Ligeia Mare, Punga Mare, and Jingpo Lacus) were masked out and an elevation threshold was applied to each individual sea based on the average SAR topography elevation value at each lakes shoreline. In the fractal algorithm described above, a point within one of the seas was rejected if the elevation was above the average shoreline value, or accepted if the elevation was below the average shoreline value. This method forms broad topographic lows where the four largest seas are located, while still maintaining the fractal nature of the surrounding topography within and surrounding the imposed basins.

We note that the evolution of a landscape subjected to a hydrological cycle through the processes of erosion and deposition of material has an impact on the resultant hydrology. Well-developed fluvial systems will transport fluids rapidly from source to sink. Erosion and deposition can result in systematic variations in the slope and fractal character of the relief within a catchment. This study is limited by the sparsely sampled topography data available to us. Although our synthetic topography matches the fractal character and overall relief of the Titan landscape, while also reproducing known basins surrounding large seas, it cannot reproduce the exact nature of Titan’s relief. However, results of landscape evolution models indicate that fluvial erosion has resulted in only slight (<9%) changes to the surface topography (Black et al.,
2012), and the density of drainage networks is generally low (Burr et al., 2013), supporting our use of a simple fractal surface. To test the effect of a more evolved Titan landscape, models were run (not shown) in which runoff was controlled by a diffused representation of the surface topography, which acts to smooth over local lows and increase the size of catchment areas.

Figure 2.3. Fractal topographic map for the basin-scale modeling, generated based on the fractal properties of the SAR topography data over the region surrounding Ligeia Mare.

Figure 2.4. The SAR topography at the north polar region used in the interpolation algorithm overlain by lakes (black) mapped from the Cassini radar map of the north pole (a) and the fractal topographic map with an interpolation algorithm and imposing Kraken Mare, Jingpo Mare, Ligeia Mare, and Punga Mare centered at $0^\circ$W longitude (b).
2.2.2 General Circulation Model

The general circulation model (GCM) used for this study (Newman et al., 2011, 2015) is a global adaptation of the terrestrial Weather, Research, and Forecasting (WRF) model described in detail by Richardson et al., (2007). In contrast to the limited area of the original WRF model, TitanWRF (and the more general PlanetWRF) uses a global atmospheric model that accounts for planetary scale physics. For Titan specifically, visible and near infrared absorption by methane, Rayleigh scattering, and haze particle scattering are accounted for. The GCM utilized for this study (Newman et al., 2015) limited liquid methane at the surface by allowing individual cells to dry out if evaporation exceeded precipitation over time, and included the latent heat effects for methane vaporization, but lacked topographic effects. The GCM output was at a spatial resolution of 5.625° longitude by 5° latitude and a temporal resolution of 128 days. While shorter time steps were investigated, because of the longer timescales for subsurface flow, higher frequency GCM outputs had little effect on the seasonal stability and distribution of Titan lakes. The GCM was initialized with uniform surface methane and then run until its methane cycle reached equilibrium – i.e., until the annual-mean surface methane distribution remained roughly constant from year to year. By this point in the GCM the low and mid latitude surface was dry except for transient surface methane immediately following precipitation events, with long-term surface methane remaining only poleward of ~75°N. After equilibration of the GCM, the Titan-yearly time series of longitudinally averaged potential evaporation \(E_p\) and precipitation \(P\) were then determined from the average of an 18 Titan-year GCM simulation and used as the inputs to the hydrology models. The potential evaporation was determined using a mass transfer equation given in general form as:

\[
E_p = C_u \Delta q
\]  

(1)
where $C_u$ is a density-weighted exchange coefficient related to the surface wind stress and $\Delta q$ is the difference between the actual and saturated specific methane humidity. Figure 2.5 shows $E_p$ and $P$ for a range of latitudes, averaging over all longitudes. Note that the large increase in $E_p$ at 75°N coincides with the boundary between a generally methane-free and methane-covered surface in the GCM. This sharp transition is due in part to the effect of evaporative cooling, which decreases the surface temperature in the polar region and reduces further evaporation.

The GCM outputs used for this work encompass hyper-arid climates below 75°N down to 45°N, with an average precipitation rate of 2.37 cm/yr and an average evaporation potential of 635 cm/yr, to semi-arid climates above 75°N, with an average precipitation rate of 5.79 cm/yr and an average evaporation potential of 11.23 cm/yr. During northern summer and spring, an increase in the solar insolation causes higher evaporation potential and highly variable precipitation rates around the pole, while northern winter predicts lower evaporation potential and precipitation rates. At latitudes > 75°N, summer and spring precipitation rates range from <1 cm/yr to 54.7 cm/yr with an average precipitation rate of 10.1 cm/yr and an average evaporation potential of 18.5 cm/yr. The winter months are characterized by low precipitation and evaporation potential, and an $E_p/P$ ratio closer to 1. Hyper-arid climates below 75°N are characterized by high evaporation potentials and little change in precipitation over the course of a Titan year.

The hydrological modeling in this work (see Section 2.2.4 below) allows methane to flow in the subsurface between latitudes and in some cases to evaporate from or even pond on the surface at lower latitudes where no stable surface methane is predicted by the GCM. However, the evaporation potential and precipitation outputs from the GCM were not coupled to the hydrological model and thus lacked the inherent feedbacks between the deep subsurface and
atmospheric components of the hydrological cycle. For example, the subsurface flow of methane to drier low latitudes, below 60°N, would result in evaporative cooling and decreased evaporation rates relative to the evaporation potential predicted by the GCM. These decreased evaporation rates could allow stable lakes to form at lower latitudes. These limitations could only be resolved with a fully coupled land and atmosphere model, which is beyond the scope of this work. Nevertheless, in simulations with realistic rates of recharge and runoff at the high polar latitudes, the low rates of subsurface flow to the lower latitudes result in evaporation rates that are much lower than the predicted evaporation potential rates and thus would be unaffected by a modest decrease in the evaporation rate due to evaporative cooling.

Figure 2.5. Precipitation rates (a) and evaporation potential (b) from the general circulation model from 45°N to 90°N over a Titan year. Evaporation potential is saturated in the panel (b) at lower latitudes to highlight the variability at high latitudes but reaches values greater than 1000 cm/yr.

2.2.3 Aquifer Recharge and Runoff Generation

Precipitation that reaches the surface can infiltrate and percolate downward to the methane table, exceed the infiltration capacity and generate runoff, evaporate from the bare soil surface,
or remain in the soil and/or unsaturated zone. These behaviors are controlled by processes at scales much smaller than the resolution of the subsurface hydrology model used in this study, and rely on poorly constrained parameters for Titan such as soil properties and soil thickness. In order to capture the nature of these processes occurring at the interface between the atmospheric and surface components of the hydrological cycle, we adopted a simple empirical method used in terrestrial hydrology. Budyko-type methods (Budyko, 1974) approximate these micro-scale processes on a basin-scale using empirical discharge data from terrestrial basins to derive a functional relationship dependent on the aridity index ($\phi$), defined as the ratio of the mean annual evaporation potential and the mean annual precipitation. Earth-based studies (McMahon et al., 2011) have shown that for basin-scale hydrology over long periods, this functional relationship takes the form:

$$F(\phi) = 1 - e^{-\phi}$$

(2)

where $F$ is the fraction of precipitation that evaporates directly from the surface and does not contribute to recharge or overland flow. Using this functional relationship, the precipitation and aridity index can be used to calculate the recharge and runoff generation as a function of time, $t$:

$$I(t) = P(t)[1 - F(\phi)]$$

(3)

where $I(t)$ is the excess precipitation that will become either surface runoff or aquifer recharge. The simple form of Eq. (2) is useful because it allows easy scaling to account for the possibility of different behavior of methane on Titan in comparison to water on Earth, as is discussed below.

Although the Budyko-type relationship in Eq. (2) generally holds for Earth, it may not apply directly to Titan. The liquid methane interaction with both the ice grains and mantling solid organics will have a controlling effect on the fraction of the precipitation that will participate in the surface and subsurface hydrology rather than evaporate directly back into the
atmosphere. To account for this uncertainty, models were also run in which the Budyko-type relationship was modified to either increase or decrease the fractional amount of precipitation excess. This variability was modeled by adding an arbitrary scaling factor to the aridity index in the exponent (Eq. (1)), with the limiting case of a scale factor of zero, allowing all precipitation to contribute to the surface and subsurface hydrology.

The excess methane precipitation (as determined by the scaled Buydko-relationship) is partitioned into methane that recharges the methane table, and methane that contributes to the surface runoff by means of shallow subsurface flow, flow through a partially saturated soil layer, and direct overland flow. On Earth, the partitioning of excess precipitation between aquifer recharge and surface runoff is dependent on the soil and subsoil layer thicknesses, the slope of the land, the total upslope area discharging to a specific region, the rate of precipitation during a rainfall event and the permeability of the soil. Depending on the properties of the soil and subsoil layers, the recharge to the aquifer can range from a negligible fraction of the excess precipitation in the case of a saturated clay layer or exposed bedrock, to nearly all of the excess precipitation in the case of a highly permeable sand layer. The permeability of soil on Earth is largely controlled by the grain size of the soil and the amount of organic material it contains. The regolith on Titan likely contain a mixture of ice grains and solid organics (Lorenz et al., 2008b), though it is unknown if these organics would play a similar role to those in soils on Earth. The Huygens probe landed in the equatorial region of Titan and found a mixture of solid organics at the surface (Niemann et al., 2005). Results from mechanical probing of the surface substrate were consistent with a tar-like material, possibly similar to a saturated soil or a lightly packed snow on Earth (Zarnecki et al., 2005). The nature of the interaction of liquid hydrocarbons with the ice and solid hydrocarbons that mantle the surface is unknown, and will have a controlling
effect on the amount of recharge to the aquifer. Due to the poor constraints on the subsurface properties, the unknown nature of the interaction of the liquid with the regolith on Titan, and the large spatial and temporal scales used in this model, recharge and runoff fractions were homogenously set over the entire domain and varied between 0 and 1 to investigate the sensitivity of the results to these parameters. Spatially and temporally varying recharge and runoff fractions were also investigated but found to have little effect on the results given the long temporal and spatial scales of interest in this study.

2.2.4 Fluid Flow through a Porous Medium

Once methane has infiltrated into the subsurface and reached the methane table, it will contribute to subsurface flow in an unconfined aquifer. Subsurface flow through a porous medium was modeled using a finite-difference approximation of the groundwater flow equation in an unconfined aquifer. The lateral flow of fluid in an aquifer depends on the active aquifer thickness \( b \), the hydraulic conductivity \( K \), the porosity \( n \) and the hydraulic gradient \( \frac{\partial h}{\partial x} \):

\[
\frac{\partial}{\partial x} \left( K_x b \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left( K_y b \frac{\partial h}{\partial y} \right) = n \frac{dh}{dt}
\]

where the active aquifer thickness \( b \) varies in space and time as a function of the hydraulic head \( h \) and the topography \( z \):

\[
b(x, y) = h(x, y, t) - z(x, y) + d
\]

in which \( d \) is the total thickness of the aquifer (assumed to be a constant value of 10 km).

The hydraulic conductivity describes the ability of a fluid to flow through a medium, and is a function of the aquifer permeability \( k \), the kinetic fluid viscosity \( \mu \), the fluid density \( \rho \), and the gravitational acceleration \( g \):

\[
K = \frac{k \rho g}{\mu}
\]
While liquid on Titan is a likely mixture of multiple constituents (Brown et al., 2008), methane and ethane are thought to be the most abundant and to control the physical properties that govern fluid flow. At the northern polar region, methane is likely the dominant fluid participating in the fluvial dissection and atmospheric component of the hydrologic cycle (Tan et al., 2013; Mitchell et al, 2015) and a pure methane fluid in the subsurface was assumed for this work. At Titan surface conditions, methane is less dense (450 kg/m³) and less viscous (2×10⁻⁴ Pa-s) than water on Earth (Lorenz et al., 2003). The low gravitational acceleration on Titan (1.35 m/s²) decreases the rate at which fluid will flow compared to flow under otherwise similar conditions on Earth. The opposing effects of the decreased density and gravity, and the decreased viscosity partially cancel out. For a given permeability, the hydraulic conductivity of an aquifer on Titan will be a factor of ~3 lower than that of an identical aquifer on Earth.

$$\frac{K_{Titan}}{K_{Earth}} \approx \frac{\rho_{methane} \mu_{Titan} \mu_{water}}{\rho_{water} \mu_{Earth} \mu_{methane}} \approx \frac{1}{3}$$  \hspace{1cm} (7)

The permeability of a porous medium is controlled by the size, shape, and interconnectivity of the pore space and fractures in the medium. Titan’s surface is composed of water ice “bedrock” overlain by “soil” that is likely composed of a mixture of ice grains and organics of atmospheric origin (Lorenz et al., 2008b), compared to silicate-based bedrock and soil composed of a mixture of silicate grains and organics of biological origin on Earth. Although the composition of Titan’s surface is different from Earth, the physical properties governing the flow through the porous media may be similar in aquifer systems. Flow through the soil layer is controlled by the grain size, and the presence of widespread sand dunes on both Earth and Titan supports similar, or at least overlapping, grain size distributions of some of the materials at the surface. Flow through the crystalline crusts of both Titan and Earth would occur dominantly through fractures, whose apertures as a function of depth are determined by the
fractal nature of the fracture surfaces. Fracture apertures on Earth follow a predictable decrease with depth irrespective of host rock lithology (Snow, 1970), allowing for generalized permeability models for Mars (Hanna and Phillips, 2005) and Titan. The reduced Young’s modulus of ice (Gammon et al., 1983) would likely lead to more rapid closure of fractures with depth, which may cancel out the effect of the lower gravity on Titan.

Permeability values can span orders of magnitude in terrestrial aquifers, and there is no reason to believe that aquifers on Titan would be fundamentally different or fall outside the range of values encountered on Earth. Here we use a range of permeability values representative of terrestrial aquifers. We assume a laterally homogenous aquifer in which the vertically averaged permeability has an exponential dependence on the depth of the methane table:

\[ k(h) = k_0 e^{-(h/z_0)} \]

where \( k_0 \) is the vertically averaged permeability when the methane table is at the surface and \( z_0 \) is a scale height which we assume to be 5 km. After scaling for effective stress \((\rho_{rock} - \rho_{fluid})g\), this scale height is equivalent to 0.2 km on Earth, which is less than the value for Earth of ~1 km from an exponential fit to the model of Manning and Ingebritsen (1999) between 1 and 5 km depth, but greater than the equivalent scale height after correcting for the smaller Young’s Modulus of ice relative to rock (~0.1 km). Permeability values on Earth commonly range between \(10^{-16}\) and \(10^{-4}\) cm\(^2\) (Heath, 1983), with typical values for unconsolidated sand of \(10^{-9}\) to \(10^{-4}\) cm\(^2\), for sandstones of \(10^{-12}\) to \(10^{-8}\) cm\(^2\), for carbonate rocks of \(10^{-10}\) to \(10^{-4}\) cm\(^2\), and for igneous aquifers of \(10^{-16}\) to \(10^{-7}\) cm\(^2\). In this study, we examine values of the vertically averaged permeability when the methane table is at the surface of \(1 \times 10^{-6}\) cm\(^2\) (corresponding to a permeability at the surface of \(k_{surf}=2.3 \times 10^{-6}\) cm\(^2\)), \(1 \times 10^{-8}\) cm\(^2\) (\(k_{surf}=2.3 \times 10^{-8}\) cm\(^2\)), and \(1 \times 10^{-10}\) cm\(^2\) (\(k_{surf}=2.3 \times 10^{-10}\) cm\(^2\)). For the purpose of discussion, we will refer to these as high (comparable to a fractured
bedrock aquifer), intermediate (comparable to an unfractured sandstone or a slightly fractured granite aquifer), and low permeability (comparable to a consolidated bedrock aquifer), respectively. Models run using permeability values an order of magnitude higher and lower than the, highest and lowest permeability values were found to have little effect on the overall results, and thus justify the choice of permeability range used in this work. As will be shown below, the results favor the intermediate permeability models, obviating the need to consider higher and lower values outside of this range. We note that this range of assumed permeability values over four orders of magnitude exceeds the likely range of viscosities of hydrocarbon mixtures in the subsurface of Titan (Lorenz et al., 2010; Hayes et al., 2013), and thus the specific fluid composition and viscosity has only a secondary effect on our conclusions.

While a laterally homogenous aquifer is assumed for this study, most aquifers on Earth are not homogenous. Studies have found that smaller scale heterogeneities within the aquifer have little influence on the regional scale flow (e.g., Freeze and Witherspoon, 1967), suggesting that heterogeneities will have little effect at the spatial scales used in this work. The use of a single average aquifer permeability throughout the entire domain allows the investigation the influence of different aquifer properties on the hydrology and formation of lakes. The potential for karst morphology on Titan (Mitchell et al., 2008; Cornet et al. 2015) suggests that focused high permeability pathways and interconnected cave systems may exist on Titan. A karst aquifer system can influence the distribution of methane and the subsurface hydrology, and may not be properly modeled as Darcy flow. However, flow in cave systems behaves more like overland flow, which is included in our model, rapidly redistributing subsurface methane. Models in which all of the precipitation contributes to runoff rather than recharge may approximate a well-
developed karst system. Further discussion of the influence of focused flow paths and hydrologic settings are found in Section 4.3.

The porosity is assumed to follow a similar exponential relationship to that assumed for the permeability

\[ n(h) = n_0 e^{-(z-h)/d} \]  

(9)

in which the same scale height of 5 km is used. On Earth, the effective elimination of pore space and permeability with depth is dominantly a result of pressure solution rather than elastic compaction \((\text{Renard et al., 2000})\). On Titan, it is unknown what processes controls the variations of permeability and porosity with depth. However, since the total range of relief in our models is typically \(<2\) km, the decrease in porosity and permeability with depth has only a minor impact on the subsurface flow and different scale heights would not significantly alter our results. We assume a maximum aquifer depth of \(d_{\text{max}}=10\) km (equivalent to a depth of \(~0.4\) km on Earth by simple effective stress scaling). However, our results will not be sensitive to this choice for any maximum aquifer depth that is substantially greater than the typical vertical relief on Titan of \(~2\) km. Furthermore, the deepest portions of the aquifer will have little effect on the total flow due to the low porosity and permeability resulting from the assumed exponential relationships.

### 2.2.5 Surface Runoff

Some fraction of the methane falling on the surface from precipitation will contribute to runoff and aquifer recharge, as determined using the Budyko-type method described above. While multiple processes may control runoff and channelized flow on Titan’s surface \((\text{Burr et al., 2013})\), the large time steps associated with numerical subsurface modeling in comparison with the timescales for dynamic surface runoff necessitate the use of a simple analytic approximations to represent runoff. In drainage basins on Earth, the precipitation excess does not
instantaneously discharge at the outlet, but will be delayed by a storage component dependent on the size and slope of the catchment. For this study, a linear reservoir approximation is used to determine surface runoff based on the assumption that the amount of liquid stored in a catchment is linearly related to the amount of surface runoff (Cunge, 1969; Overton, 1970; Dooge, 1973)

\[ S = KQ \]  

(11)

where \( S \) is the storage in the catchment in kilometers (equivalent to the average depth of methane across the catchment), \( Q \) is the surface runoff in kilometers per hour, and \( K \) is the storage coefficient in hours. Studies have found that the storage coefficient \( (K) \) is related to the time between the initial precipitation event and the peak runoff at the catchment outlet (Overton, 1970):

\[ K = \frac{t_c}{2} \]  

(12)

where \( t_c \) is the concentration time for a catchment defined as the time it takes a parcel of liquid to travel the length of a catchment. The change in catchment storage over time is determined by the balance between the excess precipitation in the basin and the surface runoff at the basin outlet

\[ I(t) - Q(t) = \frac{ds}{dt} \]  

(13)

where \( I \) is the fraction of the excess precipitation that will runoff determined using the Budyko-type method. Combining Eqs. (11) – (13) yields:

\[ I(t) - Q(t) = \frac{t_c}{2} \frac{dQ}{dt} \]  

(14)

which is solved for \( Q(t) \) using an explicit finite difference approach (Overton, 1970). The characteristic time can be determined using either an analytical solution to the kinematic wave equation for smaller watersheds or derived from empirical catchment data (Watt and Chow, 1985). A relationship between the geometry of the catchment and the time it takes a parcel of
liquid to travel the length of the catchment (the concentration time, $t_c$), was derived by Watt and Chow (1985) for large catchment basins (up to 5800 km$^2$):

$$t_c = 0.128 \left( \frac{L}{S_0^{0.5}} \right)^{0.79}$$  \hspace{1cm} (15)

where $L$ is the length from the furthest reach of the catchment to the outlet in km and $S_0$ is the dimensionless slope of the catchment. In our treatment of overland flow, we use values of $L$ and $S_0$ from individual catchments in the model topography. As discussed in Section 2.1, catchment lengths in our model topography match those determined from the mapping of channels. While the concentration time is an empirically derived value related to the slope and length of the catchment, the flow time in an open channel is inversely related to the velocity of the flow. Velocity will scale as the gravitational acceleration over the kinetic viscosity (the dynamic viscosity ($\mu$) over the density ($\rho$)), thus flow time is proportional to $\mu/\rho g$. Due to the lower gravitational acceleration ($g$) of Titan and the lower density and viscosity of liquid methane ($\mu$), the time that methane takes to traverse a catchment on Titan should be longer than the time for water to traverse an identical catchment on Earth. For Titan, we scale the concentration time for Earth as:

$$t_{c,Titan} = t_{c,Earth} \left( \frac{g_{Earth}}{g_{Titan}} \right) \left( \frac{\rho_{water}}{\rho_{methane}} \right) \left( \frac{\mu_{methane}}{\mu_{water}} \right)$$  \hspace{1cm} (16)

which yields a factor of 3.23. The treatment of overland flow results in a slight lag between the peak precipitation and peak discharge into lakes for the largest catchments. This is included in the models for the sake of completeness, but does not have a significant effect on the results.

While surface liquid was not explicitly channelized in the model, the spatial patterns of overland flow were calculated using an eight-point routing method (Wang and Hjelmfelt, 1998) in order to determine the size, geometry, and outflow pattern of basins. This method determines surface flow direction at a particular cell based on the steepest downhill slope in the eight
surrounding cells. For the catchment scale models, the fractal nature of the synthetic topography resulted in catchment basins with total catchment areas ranging from 27 km$^2$ to over 2000 km$^2$ with an average of 389 km$^2$. Lengths scales of these basins, calculated from basin mouth to divide, ranged from <10 km to 57 km with an average of 19 km. Polar scale models had total catchment areas ranging from 62 km$^2$ to 5200 km$^2$ and lengths ranging from <10 km to 115 km.

Observed fluvial dissection on Titan ranges in spatial extent from the submeter scale (Perron et al., 2006), as observed during the Huygens probe decent, to >1000 km (Burr et al., 2013). The majority of identified channels having lengths less than 50 km (Langhans et al., 2012), although this estimate is likely underestimated due to the lack of coverage. Channels measured in another study yield a mean length of ~230 km and median length of 155 km for valleys pole-ward of 70°N (Burr et al., 2013). Cartwright et al. (2011) mapped several drainage basins feeding large north polar lakes, finding a catchment area between 1760 km$^2$ and 2640 km$^2$ for at least one mapped basin.

Thus, despite the lack of channels in the synthetic fractal topography, the size distribution of drainage basins in the model is comparable to that observed on Titan. Our fractal topography lacks the long channels observed in some locations on Titan (Langhans et al., 2012; Burr et al., 2013). However, our average basin lengths reflect the entire surface surrounding the pole, while geomorphic studies are likely biased towards the most well-developed and clearly expressed valley systems. The overall low density of networks in imaged areas of the pole suggest large undissected regions, though this may in part be limited by the quality of the data (Burr et al., 2013).
2.2.6 Lake scheme

Lakes and seas on Earth have a uniform surface elevation (hydraulic head) that fluctuates in time due to influx and outflux. Earth based models (Harbaugh, 2005; Maxwell and Miller, 2005; Fan et al., 2006) typically impose large bodies of water as constant hydraulic head boundaries throughout the model run. Here we are investigating the influence of hydrological properties on the lake area distribution and seasonal lake area change on Titan, and it is important to allow lakes to form naturally and to fluctuate over time. Liquid that ponds on the surface in these models was run through a lake diffusion scheme in order to bring each lake surface to an approximately constant hydraulic head. This lake treatment solved Eq. (4) at lake pixels over multiple iterations for each iteration of the subsurface model, with the porosity of the free surface liquid set to 100% and a high permeability value of \(10^{-6}\) \(\text{cm}^2\). The lake diffusion was only applied to pixels where the methane table was above the surface. Liquid was allowed to diffuse over the surface as needed to allow natural adjustments of the shoreline.

2.3 Basin-Scale Hydrology of Titan Lakes

We first focus on a set of basin-scale models to explore the effects of the aquifer properties for identical topography and climate. The basin-scale modeling used synthetic fractal topography in a 674×674 km spatial domain (in order to encompass a region twice the size of Ligeia Mare) at a resolution of 5.225 km/pixel with multiple catchment basins feeding a central topographic low (Figure 2.4). A set of models was run to investigate the influence of aquifer permeability, with \(k_0\) ranging from \(10^{-10}\) to \(10^{-6}\) \(\text{cm}^2\), runoff and recharge fractions both set at 50%, and a surface porosity of 30%. Other model sets were designed to test the sensitivity of the lakes to the fraction of runoff and recharge, the amount of precipitation that will evaporate from the surface prior to runoff and recharge generation, and the porosity of the aquifer. All model sets
used the latitudinally-averaged GCM output between 75-90°N, for which the Budyko-type model predicts an annual precipitation excess of ~14% of the total precipitation that reaches the surface, distributed as recharge and runoff. The model results were analyzed by looking at the distribution of lakes and hydraulic head in map view, the time evolution of the different fluxes into the lakes over the entire model domain (direct precipitation, direct evaporation, overland flow, and subsurface flow), and the time evolution of the lake area as a fraction of the total area of the model domain.

### 2.3.1 Influence of the Aquifer Permeability on the Lake Hydrology

The distribution and flow of methane in the subsurface exerts a strong control over the sizes and distribution of lakes on the surface. Thus, the size-frequency distribution of lakes, the hydrological budget of the fluxes in and out of the lakes, and the rate of lake level change all depend on the properties of the aquifer. Permeability can vary over several orders of magnitude, and thus exerts the strongest control over the lake hydrology. We performed a set of simulations with identical surface runoff scaling and climate conditions, but with a permeability that varied from $10^{-6}$ cm$^2$ to $10^{-10}$ cm$^2$. A high permeability aquifer (e.g. $10^{-6}$ cm$^2$) predicts a large lake with an area of $3 \times 10^4$ km$^2$ to form in the central basin, with small ($\sim 10^2$ km$^2$) and intermediate ($\sim 10^3$ km$^2$) lakes scattered at an average distance of 225 km from the central basin where the methane table intersects local topographic lows (Figure 2.6a). The methane table in these models follows a strongly diffused representation of the surface topography, primarily intersecting the surface in the central basin and leaving much of the surrounding surface perched high above the methane table. This results in maximum lake depths of ~450 m for the large central lake and lake depths between 150-250 m for intermediate and small lakes (Figure 2.6b). The central lake depth predicted by the model is approximately 3 times greater than the maximum depth of a single
bathymetry profile across Ligeia Mare (Mastrogiuseppe et al., 2014), but does not account for sedimentation within the lake. Due to the fractal nature of the topography, the central lake does not form within a single low depression but instead within a series of small depressions interconnected by shallow necks, resulting in complex lake morphology. The purely fractal nature of the topography also has some influence on the size frequency and depth of the small lakes, which may be influenced by dissolution not represent in the model.

This model also allows us to examine the fluxes of methane into and out of the lakes over the course of a Titan year (Figure 2.6c). The volume fluxes in and out of an average lake were calculated by summing the fluxes into all lakes as functions of time and dividing by the total number of lakes. High and variable precipitation rates during the summer months ($L_s \approx 90^\circ$) results in alternating wet and dry periods. During the dry periods, high evaporation from the lake surfaces causes the lake levels to drop below the methane table allowing subsurface flow into the lakes from the surrounding aquifer (Figure 2.6c). During times of heavy rainfall and runoff, subtle increases in the lake level relative to the surrounding methane table occur, resulting in periods of net subsurface flow to the surrounding aquifer for the small and intermediate sized lakes. This response from the aquifer mitigates the net liquid lost or gained by the lake from evaporation and precipitation at any given time. The model predicts <1% change in lake area as a fraction of the total lake area over a Titan year, which is within the limit of the model resolution (Figure 2.6d). The minimal changes in lake area are due to a combination of the approximate balance of fluxes into and out of the lakes, and the large volume of the lakes relative to the fluxes involved. The winter season ($L_s \approx 270^\circ$) is characterized by persistent low precipitation and runoff rates as well as persistently low rates of subsurface flow, with the latter providing the dominant flux into the lake. Lake level change is minimal during this period and subsurface flow is
Figure 2.6. High permeability model ($10^{-6} \text{ cm}^2$) showing the relative hydraulic head with lakes overlain in black (a), the lake depths (b), the average fluxes into (positive precipitation, runoff, and subsurface flow) and out of (positive evaporation and negative subsurface flow) a typical lake over a Titan year (c), and the time evolution of the total lake area as a fraction of the area of the model domain (d). The time evolution in (c) and (d) encompass on Titan year (29.5 Earth years), with spring equinox ($L_s=0^\circ$) at year 0.

...dominated by the central low hydraulic head. The high permeability of the aquifer results in subsurface flow contributing 65.2 km$^3$/Titan-yr into all lakes in the model domain or 32% of the total inflow (with 17% being sourced from runoff and 51% from direct precipitation onto the lake surfaces).

Similar to the high permeability case, the hydrology of an intermediate permeability aquifer ($10^{-8} \text{ cm}^2$) is dominated by a central lake ($\sim 2 \times 10^4 \text{ km}^2$) located in the largest topographic...
depression (Figure 2.7a) with a lake depth of ~400 m (Figure 2.7b). Unlike the high permeability case, at intermediate permeability there is not an instantaneous balance between subsurface flow and the net flux from evaporation, precipitation, and overland flow. Subsurface flow maintains a relatively constant flux into the lakes, while the other fluxes exhibit strong seasonality (Figure 2.7c). Nevertheless, there is little seasonal change in lake area since the large sizes of the lakes stabilize them against the comparatively small seasonal variations in net flux (Figure 2.7d).

Subsurface flow constitutes the primary flux into lakes during the quiescent winter months. Over the course of a year, subsurface flow accounts for 55.8 km$^3$/Titan-yr into all lakes in the model domain or 29% of the total inflow with direct precipitation and runoff accounting for 51% and 20% respectively (Figure 2.7c). In contrast to the high permeability case, the intermediate permeability results in an increase in the hydraulic head gradient, forming a greater number of small to intermediate-sized lakes where the methane table intersects the surface at small topographic depressions, while decreasing the size of the large central lake. The total lake area is somewhat reduced relative to the high permeability case, largely as a result of evaporation from the methane table where it approaches the surface, which occurs more widely at lower permeability.

In contrast to the previous models, low permeability aquifers (10$^{-10}$ cm$^2$) constrict flow through the aquifer raising the methane table close to the surface throughout the model domain. This prevents large lakes from forming in the deepest basin and limits the size of the largest lake to 850 km$^2$ (Figure 2.8a), which is only ~3% of the size of the largest lake in the high permeability model. Small lakes are now favored within small local topographic basins throughout the model domain with an average lake depth of ~40 m (Figure 2.8b). The predicted subsurface flux shows little change over the course of a Titan year and is dominated by short
Figure 2.7. Intermediate permeability model ($10^{-8} \text{ cm}^2$) showing the relative hydraulic head with lakes overlain in black (a), the lake depths (b), the average fluxes into and out of a typical lake over a Titan year (c), and the time evolution of the total lake area as a fraction of the area of the model domain (d).

flow paths to local topographic lows. The subsurface flux into lakes is reduced relative to the high permeability case (27.2 $\text{km}^3$/Titan-yr or 16% of the total inflow to the lakes), increasing the fractional input from runoff (33% of the total inflow) and direct precipitation (51% of the total inflow) (Figure 2.8c). The contribution from subsurface flow does not scale directly with the permeability because of the greater hydraulic gradients at low permeability. The smaller size of the lakes compared to the high and intermediate permeability cases results in greater sensitivity
of the lakes to variations in evaporation and precipitation, and larger seasonal changes in lake area. The total fractional lake area ranges from 10.8% to 12.8% of the total model area (Figure 2.8d), amounting to a ~16% variation in the lake area as a fraction of the total lake area over a Titan year. A large increase in lake area is predicted to occur in the mid to late northern winter as the evaporation rate from the lake surface decreases.
2.3.2 Sensitivity to other parameters

We examined a wide range of parameter space in all of the terms that affect the hydrology, finding that most have only a second-order effect on the results. The assumed porosity controls the storativity of the aquifer, and thus the time-dependent response of the aquifer to any applied forcing. Substantially reduced porosity (5% surface porosity) results in a larger seasonal change in lake area, but does not otherwise change the model predictions.

The assumed Budyko relationship determines what fraction of the precipitation is available for hydrological activity as either runoff or recharge, rather than directly evaporating from the surface. However, the partitioning of this fluid into runoff and recharge is poorly constrained given the unknown nature of Titan’s surface materials and the poorly constrained interaction of liquid methane with the ice-organic mixture that likely comprises the soil. The amount of runoff compared to recharge to the aquifer affects the timing of lake level change over a Titan year, thus influencing the seasonal changes in lake area. Surface runoff acts as a rapid mechanism for influx into the lakes that closely follows the temporal distribution of precipitation, while recharge to the aquifer determines the volume of methane available for slow influx to the lakes by subsurface flow. For intermediate to high permeability aquifers, increasing either the runoff or recharge fractions to 100% of the excess precipitation has little effect on the results due to the high rates of subsurface flow in these aquifers. For the case of 100% runoff, the runoff ponds in the closest topographic depression where it infiltrates and contributes to subsurface flow, resulting in only a minor redistribution of surface fluid prior to infiltration. For low permeability aquifers, increasing the recharge fraction to 100% of the excess precipitation results in a decrease in the total lake area and a large seasonal lake area change. In this case, the aquifer recharge is inefficiently removed by subsurface flow resulting in a rise in the methane
table toward the surface, where methane can be lost to evaporation. As a result of the increased evaporation from the methane table, the total lake area is reduced.

There is also significant uncertainty in the fraction of the precipitation that will evaporate directly from the surface prior to generating runoff or aquifer recharge (the applied Budyko-relationship). The Earth-based Budyko relationship allows 14% of the precipitation to generate runoff and recharge. The lack of plant cover on Titan in comparison to Earth may reduce the tendency to store liquid in the near-surface soil layer, and thus reduce the fraction of the precipitation that is subject to direct evaporation from the soil layer. We tested this with a model in which no Budyko scaling was applied, thus allowing all precipitation to generate either runoff or aquifer recharge. This has the effect of significantly increasing the lake area for all models. Scaling the Budyko relationship to allow ~46% of the GCM predicted precipitation to either runoff or recharge produces a Ligeia Mare-sized lake for a high permeability aquifer, suggesting that the Earth-based relationship may underestimate recharge and runoff on Titan. However, this does not alter the fundamental conclusions regarding the role of permeability in controlling the balance between small to intermediate size lakes and large seas.

2.4 Hydrology of the North Polar Lake Region

The previous basin-scale hydrological models described above gave insight into the general behavior of hydrology on Titan, but focused on an isolated basin influenced by an average GCM output over the latitudes at which lakes are observed. Large basins on Titan can span a range of latitudes (e.g. Kraken Mare) and may be influenced by differing climates at different latitudes. Furthermore, flow both within the lakes (Lorenz, 2014) and in the subsurface from more temperate high latitudes to more arid low latitudes may significantly affect the hydrology. By using the GCM output to spatially vary the precipitation and evaporation potential
values over a larger model domain representing the entire north polar region, I was able to investigate the importance of the latitudinal variation in climate and long-distance subsurface latitudinal transport on the latitudinal distribution of lakes (Figure 2.9a). The polar model was also used to investigate the effects of decreased evaporation over large polar seas due to either accumulated ethane or other solutes (Lorenz, 2014) or to lake effect changes to the local climate (Tokano, 2009). This was implemented in the hydrological model by simply decreasing the evaporation potential over Kraken Mare and increasing the total amount of aquifer recharge and surface runoff at the high polar latitudes. The possible effects of a non-uniform permeability distribution at the north polar region was also investigated by varying the permeability with latitude as supported by evidence for karst at high polar latitudes (Mitchell et al., 2008; Cornet et al., 2015).

The polar model encompassed the north polar region down to 45ºN, in a 4000 km × 4000 km domain (Figure 2.9b). The model used a flat circular domain rather than a curved spherical cap for simplicity. This has the effect of slightly overestimating the divergence of subsurface flow from high to low latitudes, and thus slightly underestimating the local flux at lower polar latitudes. However, since the hydrological processes of interest dominantly occur at latitudes above 75ºN and at the transition to more arid climates at 75ºN, this simplified geometry has little effect. For the simple case of a constant volumetric flow of methane from high to low latitudes, the assumed flat model domain decreases the local flux per unit distance crossing 75ºN by only 1% relative to what a spherical cap model would predict. This model used synthetic topography generated from a modified fractal scheme so as to capture the long wavelength nature of the topography around the pole and impose basins for the largest seas, as discussed in Section 2.1. With this topographic model, the conditions needed to form both the large seas, in particular
Kraken Mare, and the distribution of smaller lakes can be investigated. The GCM inputs to the hydrological model were spatially distributed in 5º latitude increments in a polar geometry. The lake area as a function of latitude was compared to the observed lake area distribution over the same north polar region with and without Titan’s largest sea, Kraken Mare (Figure 2.9a). The removal of Kraken Mare from the observed lake distribution allows for the comparison of the distribution of the smaller lakes with the distribution predicted by the models, as Kraken Mare dominates the north polar lake area.

2.4.1 Influence of Aquifer Properties on the Latitudinal Distribution of Lakes

Based on the results of the basin-scale models, we here consider only the high and intermediate permeability cases. The low permeability models at the basin-scale predicted many small lakes perched at high elevations, and large seasonal variation in lake area, both in conflict with observations. A high permeability aquifer (10⁻⁶ cm²) throughout the model domain allows significant subsurface flow from the pole to mid-latitudes due to a precipitation-induced latitudinal hydraulic head gradient. High precipitation rates at latitudes >75ºN generate aquifer recharge and an elevated hydraulic head relative to the more arid lower polar latitudes (Figure 2.10a), resulting in a net loss of methane from the higher latitudes by subsurface flow to low latitude sinks. The north polar hydrology in this high permeability case is controlled by topographic depressions at low latitudes, primarily the Kraken Mare basin and a low topographic region located at 50ºN in the SAR topography (Figure 2.5a) that forms a basin between 40ºN
Figure 2.9. The distribution of the fractional lake area (expressed as a percentage of the total surface area) as a function of latitude observed at the north polar region (a), with (solid) and without (dashed) Kraken Mare. The observed lake distribution was calculated from a SAR image mosaic over the north polar region down to 50°N (b; image credit: NASA/JPL-Caltech/ASI/USGS, http://photojournal.jpl.nasa.gov/catalog/PIA17655).
and 50°N in the synthetic fractal topography (Figure 2.5b). This topographic low is observed in the SAR topography and is associated with the Ganesha Macula region. While initial studies suggested a cryovolcanic origin for Ganesha Macula (Lopes et al., 2007), recent studies have found that the low topography and morphology are more consistent with an origin through hydrological and fluvial processes (Le Gall et al., 2010; Lopes et al., 2013). While some lake formation is predicted below 75°N, the high aridity at lower polar latitudes only allows a few small lakes to form and the total lake area at all latitudes predicted by this model is significantly less than the observed lake area (Figure 2.11a). The low total lake area in the north polar region, predicted by the high permeability model, is a result of the efficient movement of methane out of the wet high latitudes to the arid regions at lower polar latitudes. This efficient equator-ward transport of methane results in a nearly flat methane table, with only ~150 m variation in hydraulic head from the pole to low latitudes, leading to a methane table deep beneath the surface over most of the model domain.

Figure 2.10. Hydraulic head maps with lakes overlain in black for the polar models with a permeability of $10^{-6}$ cm$^2$ (a) and $10^{-8}$ cm$^2$ (b).
At intermediate permeability ($10^{-8}$ cm$^2$), comparable to an unfractured sandstone or a fractured granite aquifer on Earth, the polar model predicts that the vast majority of methane ponded on the surface would be contained at latitudes >75°N (Figure 2.10b). High rates of precipitation at latitudes >75°N again cause a climate-induced hydraulic head gradient from the pole to the mid-latitudes, driving equator-ward flow in the subsurface. However, the permeability in this case is too low to allow sufficient subsurface flow to permit substantial lake formation at more arid latitudes below 75°N. At higher latitudes, the increased hydraulic head and reduced subsurface flow towards the mid-latitudes results in the formation of small lakes and seas within the Ligeia Mare and Punga Mare basins, but the model does not produce large seas comparable to those observed. Although inter-latitudinal flow allows some lakes to form at lower polar latitudes in the Kraken Mare basin, the predicted latitudinal distribution of lakes for this intermediate permeability aquifer is strongly focused around the pole and does not match the latitudinal distribution of lakes below 75°N, even if Kraken Mare is excluded from the observed distribution (Figure 2.11a).

2.4.2 Influence of Increased Runoff and Aquifer Recharge

Both the high and intermediate permeability polar models fail to match the lake area distribution and to generate a Kraken-sized sea due to the high evaporation potential at the lower latitudes reaches of the Kraken Mare basin. This suggests that the Budyko-type method used in the models may be underestimating the aquifer recharge and runoff generated by the precipitation. While the observed manifestations of hydrologic activity on Titan such as fluvial
Figure 2.11. Predicted lake area as a function of latitude showing the effect of permeability (a), increased runoff and recharge (b), applying a factor to decrease the evaporation over Kraken Mare (c), and increasing the runoff and recharge while also scaling the evaporation over Kraken Mare (d).

dissection, storms, and lake morphology appear similar to their counterparts on Earth, both the solid and liquid materials involved in the hydrology are very different. Based on the complex nature of hydrocarbons in the atmosphere, at the surface and potentially in the subsurface, one might expect significant departures from processes on Earth. Particularly, the interaction of liquid hydrocarbons with solid hydrocarbons and ice is poorly understood, and may be a key process in determining the amount of methane that recharges the aquifer or generates runoff,
rather than evaporating immediately back into the atmosphere. Increased runoff and aquifer recharge might arise due to either the effects of an organic-rich “regolith” layer on Titan or enhanced recharge due to the wetting behavior of liquid methane on crystalline ice (Sotin et al., 2009).

A model with no Budyko-scaling was used in order to test an endmember scenario in which all of the methane that falls on the surface will either recharge the aquifer or runoff over the surface, with no direct evaporation of precipitation back into the atmosphere. The increased recharge to the aquifer and surface runoff raises the methane table at the high latitude polar regions, filling the Ligeia Mare and Punga Mare basins while equatorward subsurface flow reaches the Kraken Mare basin forming a small sea. As found in the previous high permeability case, the observed lake distribution is strongly underpredicted in this model (Figure 2.11b) and a Kraken-sized sea cannot form due to the high evaporation potential over the Kraken Mare basin. At the high polar latitudes, the lake area agrees with the observed lake area due to the filling of Punga Mare and Ligeia Mare, but the abundant small lakes observed extending down to 70°N are not predicted.

For the intermediate permeability case, increased recharge and runoff increases the total lake area at high polar latitudes, but inefficient subsurface flow still prevents substantial lake formation below 75°N (Figure 2.12b). Lake formation below 75°N is confined to the Kraken Mare and Jingpo Mare basins, though only small lakes form in these basins. The high polar latitudes are nearly saturated in this model, over-filling Ligeia and Punga Mare, and forming numerous other intermediate to large sized lakes that cover the majority of the high polar latitudes down to 75°N. The latitudinal distribution of lakes is significantly overpredicted at high polar latitudes and underpredicted at lower polar latitudes (Figure 2.11b). Thus, both models
without Budyko-scaling, allowing all of the precipitation predicted by the GCM to generate runoff and recharge to the aquifer, still fail to predict the observed latitudinal lake distribution and to form a Kraken-sized sea within the imposed Kraken Mare basin.

Figure 2.12. Results from the polar models showing hydraulic head maps with lakes overlain in black for a permeability of $10^{-6}$ cm$^2$ (a) and $10^{-8}$ cm$^2$ (b) with increased runoff and recharge.

### 2.4.3 Influence of Decreased Evaporation over Kraken Mare

The previous models showed that hyper-arid conditions at latitudes below 75°N prevent a large Kraken-sized sea from forming. One possible solution to this problem is that the evaporation rate over Kraken Mare may be lower than the values predicted by the general circulation model. At temperatures on Titan, the volatility of ethane is much lower than that of methane due to low ethane saturation pressure (Lunine et al., 1983; Aharonson et al. 2009), suggesting that the evaporation of ethane is insignificant on seasonal Titan timescales. Tokano (2009) found that 83% of ethane or other organic solutes within large seas at Titan’s north polar regions will inhibit the evaporation of methane, while seas with greater amounts of methane will contribute to elevated atmospheric methane humidity and fluctuating evaporation rates. These results suggest that higher concentrations of ethane or other organic solutes within the large seas
will greatly reduce the evaporation rate of methane. This effect was used to estimate the amount of ethane or organic solutes within Kraken Mare based on a mass balance within the Ligeia and Kraken Mare basins (Lorenz, 2014). Using a direct scaling of the evaporation with the organic solute concentration, that study found the involatile concentration within Kraken Mare to be ~60%, reaching greater values at lower latitudes. Other work has suggested that the composition of Titan’s largest seas are close to vapor equilibrium with the atmosphere (Mastrogiuseppe et al., 2014), which will also act to greatly reducing the evaporation from the large seas. Evaporative cooling over large bodies of liquid will also suppress the evaporation rate over Kraken Mare. The evaporation potential output by the GCM used in this work predicted high rates of evaporation potential at lower polar latitudes. However, the lack of surface methane at lower polar latitudes in the GCM produces a higher evaporation rate than if surface methane were present, due to latent heat effects as evaporation occurs. Here, we tested this effect by applying a constant scaling factor to the evaporation potential over lakes and seas within the Kraken Mare basin, with scaling factors ranging from 0.01 to 0.2. This scaling tests the net effect of evaporation suppression due to both a lake effect on the climate and the effects of the concentration of ethane or organic solutes within Kraken Mare.

We also tested a variable scaling of the evaporation rate over Kraken Mare, based on the model-predicted changes in the methane-solute ratio within the sea. In this approach, we initialized the model with a Kraken Mare with an assumed involatile volume, and let the involatile fraction evolve with time in response to the fluxes of methane into and out of the lake. These models generated similar results to the models with constant scaling of the evaporation over Kraken, and thus we focus our discussion below on the simpler models with the constant scaling factors.
At high permeabilities (10^{-6} \text{ cm}^2), a decrease in the evaporation potential over Kraken Mare to 1% of the evaporation rate predicted by the GCM results in a larger lake forming within the Kraken basin relative to the previous high permeability case (Figure 2.13a), but the total lake area and the size of the lake in the Kraken Mare basin are still much smaller than the observed lake areas (Figure 2.11c). In this model, the high rate of subsurface flow in a high permeability aquifer flushes liquid out of the Kraken Mare basin toward other small but low elevation depressions in the arid low latitudes. Similar to the previous high permeability case, the rapid removal of methane from the polar region inhibits lake formation at the higher polar latitudes, thus the primary methane sink is evaporation from small lakes and saturated surfaces at the dryer low latitudes. In this model, the suppression of evaporation over the Kraken Mare basin reduces its role in removing methane from the system to balance the precipitation and aquifer recharge at the high polar latitudes. A topographic low at a separate low latitude basin between 40°N and 50°N instead becomes the dominant sink for the system and controls the flow of methane in the subsurface (Figure 2.13a).

A high permeability model with an evaporation scaling factor of 1% over Kraken Mare, while also allowing all precipitation to either runoff or recharge the aquifer (i.e., no Budyko scaling of the precipitation), produces a lake distribution similar to the observed lake distribution (Figure 2.11d) and allows a Kraken-sized sea to form (Figure 2.13b). This model is dominated by large seas in the north polar regions filling the Ligeia Mare, Punga Mare, Jingpo Lacus, and Kraken Mare basins, but does not predict the large number of smaller lakes that are observed at other longitudes. While this model does fit with the observed latitudinal lake distribution, the requirement for all of the precipitation to runoff or recharge the aquifer is unrealistic, and the lack of small lakes is in contrast with the observed lake distribution making this model less than
ideal. When the evaporation over Kraken Mare is scaled by a factor of 0.1 (Figure 2.13c) or 0.4 (Figure 2.13d) in models with increased runoff and recharge, a lake smaller than Kraken Mare is predicted to form within the Kraken Mare basin. This suggests that, even with increased recharge and runoff in the system, a small evaporation scaling factor (~1%) is needed to form a Kraken-sized sea at high permeability values, and the models still fail to predict the distribution of smaller lakes.

In the case of the intermediate permeability aquifer with a 1% evaporation scaling over Kraken Mare, a large Kraken-sized sea forms in the Kraken Mare basin (Figure 2.14a). The latitudinal lake distribution predicted by this model agrees with the observed lake distribution (Figure 2.14b). This model predicts Ligeia Mare and Punga Mare-sized lakes at the high polar latitudes, and smaller lakes down to 75°N, but lacks small lakes between 70°N and 75°N and fails to predict a fully filled Jingpo Lacus basin. Fluid sourced from subsurface flow, direct precipitation, and runoff from the high polar latitudes remains in the Kraken Mare basin as the decreased rate of subsurface flow from the basin relative to the high permeability case and the reduced evaporation inefficiently remove fluid from the lake (Figure 2.14c). Subsurface flow acts as a net sink from Kraken Mare as a whole, though this reflects the balance of flow into the lake at high latitudes and away from the lake at low latitudes. With the exception of Kraken Mare, dry conditions below 75°N and a lack of subsurface flow prevents stable lake formation at low latitudes. At the lowest topographic depressions below 75°N, the methane table lies incident with the surface, but the high evaporation potential at these latitudes prevents methane from ponding.
Figure 2.13. Polar model results showing hydraulic head maps for a high permeability case with 1% evaporation scaling over Kraken Mare (a), 1% evaporation scaling over Kraken Mare and increased runoff and recharge (b), 10% evaporation scaling over Kraken Mare and increased runoff and recharge (c), and 40% evaporation scaling over Kraken Mare and increased runoff and recharge (d).
Similar to the previous high permeability model, any increase in the evaporation potential over Kraken Mare prevents a Kraken-sized sea from forming (Figure 2.15a). Intermediate permeability models with high recharge and runoff and applying 1% scaling to the evaporation potential over Kraken Mare greatly overestimate the abundance of lakes at high latitudes (results not shown). Models with high recharge and runoff with 5%, 10%, and 20% scaling of the
evaporation potential over Kraken Mare also overpredict lake formation at the high polar latitudes and fail to form a sufficiently large sea within the Kraken Mare basin (Figure 2.15b-d).

Figure 2.15. Polar model results showing hydraulic head maps for an intermediate permeability case with 5% evaporation scaling over Kraken Mare (a), 5% evaporation scaling over Kraken Mare and increased runoff and recharge (b), 10% evaporation scaling over Kraken Mare and increased runoff and recharge (c), and 20% evaporation scaling over Kraken Mare and increased runoff and recharge (d).

2.4.4 Influence of a Non-Uniform Permeability Distribution around the North Pole

The models above assumed a laterally uniform permeability distribution in the model domain but this may not be the case at the north polar region. Recent observations of the north polar region with Cassini’s Imaging Science Subsystem (ISS) found a bright region at ISS
wavelengths encompassing the northern lakes with a drop off in albedo at lower latitudes (Turtle et al., 2013). This suggests that the surface material at the north polar region differs from surface material at the mid-latitudes. The aquifers in the north polar region may contain soluble organic fallout material (Lavvas et al., 2008; Krasnopolsky, 2009). The presence of morphologies consistent with karst features near the polar regions (Mitchell et al., 2008) suggests that the permeability at the north polar region may be higher than that over the majority of Titan as a result of dissolution in the subsurface (Malaska and Hodyss, 2014). For karst systems on Earth, the permeability and subsurface conduit system is controlled by the regional climate, with wetter climates forming larger interconnected cave systems and drier climates inhibiting cave formation (Webb and James, 2006). The majority of rainfall that reaches the surface on Titan occurs at latitudes above 75°N, which may contribute to a more developed, higher permeability aquifer around the north polar region. Small-scale conduits and fractures in poorly developed karst aquifers will increase the average permeability of the aquifer but can still be approximated by Darcy flow in the subsurface. In contrast, large cave systems found in mature carbonate aquifers on Earth act as subsurface fluvial systems with dimensions comparable to the individual catchment basins on Titan (e.g., Mammoth cave, in south central Kentucky, has a total length of cave passages of over 600 km in a catchment with an area of ~200 km²). Flow through such a large-scale cave system will behave similar to the overland flow and not affect the model results significantly. However, the overall increase in permeability in karstic aquifers may affect flow in the north polar region.

In order to approximate variations in permeability due to karst at the north polar region, we vary permeability from higher permeability ($10^{-7}$ cm²) between 75° and 90°N to an intermediate permeability ($10^{-8}$ cm²) below 75°N. Higher permeability around the pole should result in a
better match to the sizes and distributions of high latitudes lakes, while the intermediate permeability at low latitudes will prevent rapid southward flow and loss of methane at arid low latitudes. Based on the results of a parameter sensitivity investigation, a scaled Budyko-type method was used, allowing approximately three times as much methane to generate runoff and recharge compared to the nominal models. The evaporation potential predicted by the GCM over Kraken Mare was scaled by a factor of 0.01. While other non-uniform permeability models with different choices for the permeabilities, recharge and runoff fractions, and evaporation decrease over Kraken Mare were investigated (results not shown), this model was chosen based on a qualitative and quantitative comparison with the observed lake distribution at the north polar region.

For the non-uniform permeability model, the lower permeability below 75°N restricts flow to the arid lower polar latitudes, causing increased hydraulic head above 75°N relative to the uniform permeability models (Figure 2.16a). This allows more lakes to form at latitudes above 75°N as liquid encroaches the surface at higher elevation, forming small lakes at the high polar latitudes down to 70°N and filling the Ligeia Mare and Punga Mare basins (Figure 2.16b). A Kraken-sized sea forms in that basin due to the evaporative scaling and the intermediate permeability aquifer surrounding the arid low latitudes regions of the basin, preventing the rapid removal of liquid in Kraken Mare to the subsurface. Nevertheless, there is a net loss of fluid from Kraken Mare due to subsurface flow throughout the Titan year. This is shown in the hydrograph for Kraken Mare (Figure 2.16c), which predicts substantial surface runoff and precipitation from the high polar latitudes, and net removal of methane from evaporation and subsurface flow to the surrounding low latitude aquifer. This subsurface flow from Kraken Mare likely contributes to the hydrology of smaller surrounding lakes.
Figure 2.16. The hydraulic head map (a) for the polar cap model with a 1% evaporation scaling factor over Kraken Mare and allowing 30% of the precipitation to reach the surface/subsurface hydrology. The lake distribution compared to the observed lake distribution (b) and the hydrology of Kraken Mare is shown (c).

2.5 Discussion

2.5.1 Basin-scale hydrology

For all permeability values, lakes at Titan’s North Pole are predicted to remain stable over a Titan year. While lower permeability aquifers predict larger seasonal changes and the complete...
drying of some smaller lakes (<100 km), large lakes are difficult to dry out as a result of both their large volume and their connection to the subsurface aquifers. The large sizes and lack of observed shoreline change at Titan's north polar region supports the interpretation that stable lakes are connected to a high or intermediate permeability aquifer. In contrast, the abundance of dry lakebed features (Hayes et al., 2008) supports the importance of long-term changes in the climate (Aharonson et al., 2009). The intermediate and high permeability aquifer models predict that subsurface flow into lakes comprises ~29-32% of the total fluid flux into the lakes, which is ~2 times that at lower permeability. For all permeabilities, the large contribution of direct precipitation to the lakes (~51% of the total influx) is a result of the arid climate and the assumed Budyko relationship that governs the amount of direct evaporation of the precipitation that falls on the land surface. While low permeability models predict lake areas <10^3 km^2, high and intermediate permeability models predict a large sea intermediate in size between Jingpo Lacus (1.7×10^4 km^2) and Punga Mare (4.0×10^4 km^2) but fail to produce a Ligeia Mare-sized sea. The better agreement between the size and stability of lakes predicted by the high and intermediate permeability models and the observed lakes suggests that a significant component of subsurface flow contributes to the hydrology on Titan. The extreme limit of zero permeability would be equivalent to the case in which there is no subsurface component to Titan’s hydrological cycle and the formation and stability of lakes is driven exclusively by runoff, direct precipitation, and evaporation at the lake surface. Thus, the fact that the low permeability models do not match the observed sizes and stability of lakes effectively rules out the scenario in which there is no subsurface component to Titan’s hydrological cycle.
2.5.2 Polar hydrology

Polar models at both high and intermediate permeability require a dramatically decreased evaporation potential (≈1% of the rate predicted by the GCM) over Kraken Mare in order to form a sufficiently large sea. This is easily understood as a consequence of the dramatic increase in evaporation rates below 75°N in the southernmost reaches of the basin, which exceeds the input to the basin from subsurface flow, runoff, and direct precipitation from the higher latitudes. The decreased evaporation over Kraken Mare can be brought about by the combination of an increase in the concentration of ethane or organic solutes within Kraken Mare (Tokano, 2009; Lorenz, 2014), and a lake effect on the climate over a large sea resulting in the suppression of evaporation by evaporative cooling (Tokano, 2009). These effects are important to the stability of Kraken Mare at lower latitudes. While models with an increased aquifer recharge and runoff and smaller evaporation suppression can result in a Kraken-sized sea, these models fail to fit the observed lake distribution outside of Kraken Mare.

Furthermore, models with a high permeability aquifer at the north polar region result in the rapid removal of liquid from the high polar latitudes to the dry lower polar latitudes resulting in a lower total lake area at the north polar region even when evaporative suppression is accounted for over Kraken Mare. While basin-scale models necessitate a high to intermediate permeability in order to stabilize large lakes, at long distance scales, higher permeability values can, impede lake formation at the high polar latitudes and within Kraken Mare. Models with an intermediate permeability aquifer or a high permeability aquifer surrounded by an intermediate permeability aquifer with evaporation suppression over Kraken Mare provide general agreement with the observed lake distribution at the north polar region. This suggests that while the higher polar latitudes may be characterized by dissolution in the subsurface leading to enhanced
subsurface flow, similar processes may not be at work at the lower polar latitudes where an intermediate to low permeability aquifer is required to stabilize Kraken Mare and the large seas in the arctic lake district. More importantly, we have shown that by comparing hydrological models to the observed distribution of lakes, we can shed light on the nature of the full hydrological cycle, from the subsurface to the atmosphere.

2.6 Summary and Conclusions

Titan is the only body, besides the Earth, with an active hydrological cycle. Cassini observations reveal an abundance of hydrocarbon lakes and seas in the north polar region (Stofan et al., 2007) resulting from the more temperate polar climate in comparison to the hyper-arid lower latitudes. By analogy with Earth and Mars, it appears likely that subsurface flow may feature prominently in the hydrology of Titan. Although we have very few constraints on the properties of Titan’s surface and subsurface that govern the transport of methane across and beneath the surface, the distribution of lakes provides an important observational constraint for comparison to models, which in turn can provide insight into the unseen details of Titan’s subsurface hydrology.

Basin-scale models show that high latitude Titan lakes connected to an unconfined aquifer with high ($\sim 10^{-6}$ cm$^2$) to intermediate ($\sim 10^{-8}$ cm$^2$) permeability are predicted to remain stable on the surface over a Titan year in a semi-arid climate similar to polar regions of Titan, and show little to no seasonal shoreline change, consistent with observations. Lakes of a range of sizes up to and exceeding $10^4$ km$^2$ form for these permeabilities, in agreement with the observed sizes of lakes. In contrast, lower permeability values ($\sim 10^{-10}$ cm$^2$) predict substantial seasonal changes in lake area and an abundance of smaller perched lakes rather than a mix of small lakes
Table 2.1. Summary of results and implications for the basin-scale and polar models.

<table>
<thead>
<tr>
<th></th>
<th>Model results</th>
<th>Comparison with observations</th>
<th>Implications</th>
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<tr>
<td><strong>Basin-scale models</strong></td>
<td></td>
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<tr>
<td>Low permeability</td>
<td>Large seasonal change in lake area and small scattered lakes perched at high elevations</td>
<td>Large north polar lakes and seas are observed to be stable with no discernable shoreline changes during the Cassini mission (Hayes et al., 2011)</td>
<td>Models favor subsurface flow in a high to intermediate permeability aquifer and show the importance of subsurface flow on Titan lakes and seas</td>
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<td>(10⁻¹⁰ cm²)</td>
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<tr>
<td>Intermediate permeability</td>
<td>Stable, large lakes filling topographic depressions due to the balance of evaporation and subsurface flow</td>
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<td>(10⁻⁸ cm²)</td>
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<td>High permeability</td>
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<td>(10⁻⁶ cm²)</td>
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<tr>
<td><strong>Polar models</strong></td>
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<tr>
<td>High permeability</td>
<td>High rates of subsurface flow results in the rapid removal liquid from the system</td>
<td>Total lake area is significantly lower than the observed lake distribution without methane in the hydrological system and evaporation suppression over Kraken Mare</td>
<td>Needs unrealistically high aquifer recharge and runoff with evaporation suppression over Kraken Mare to form a Kraken Mare-sized sea</td>
</tr>
<tr>
<td>Intermediate permeability</td>
<td>Decreased subsurface flow results in a raised methane table and prevents removal of liquid from Kraken Mare</td>
<td>Fits the observed distribution but fails to form a Kraken Mare-sized lake without evaporation suppression</td>
<td>Intermediate permeability and a high permeability cap surrounded by an intermediate permeability models agrees with the observed lake distribution and form a Kraken Mare-sized sea when evaporation suppression is included</td>
</tr>
<tr>
<td>Polar cap (high permeability surrounded by an intermediate permeability)</td>
<td>Diffused methane table at the high polar latitudes but flow to lower latitudes impeded by intermediate permeability aquifer surrounding the pole</td>
<td>Fits the observed distribution when a slight increase in runoff and recharge is included but fails to form a Kraken Mare-sized lake without evaporation suppression</td>
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<tr>
<td>Low permeability</td>
<td>Models not run based on results from previous basin-scale model</td>
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and large seas, in conflict with the observations. Thus, these basin-scale models support the importance of subsurface flow through intermediate to high permeability aquifers in Titan’s hydrological cycle.

Large-scale polar models with latitudinally varying climates and topography with imposed basins allow for a more direct comparison of the model results with observations for the distribution of lakes with latitude and the formation of the large seas. Intermediate permeability models predict abundant lakes at high polar latitudes, but a deficit of lakes at lower polar latitudes. High permeability models predict a deficit of high latitude lakes, but do allow a few lakes to form at lower polar latitudes due to subsurface flow from temperate high polar latitudes to hyper-arid low latitude topographic depressions. Increased recharge and runoff result in an increase in the total lake area and partially fill the larger basins in the high polar latitudes. However, all of these models both fail to produce a Kraken-sized sea and disagree with the observed latitudinal lake distribution.

Observational evidence indicates high latitudes are characterized by material of a different composition at the surface, as well as putative karstic features that are likely associated with increased permeability. Inclusion of a higher permeability polar cap surrounded by an intermediate permeability aquifer at lower polar latitudes raises the methane table and allows subsurface flow at the higher polar latitudes while preventing appreciable loss of methane to the low latitudes as well as the aquifer surrounding Kraken Mare. In this model, the combination of a somewhat increased recharge and runoff fraction and a permeability drop at lower polar latitudes results in a good agreement between the predicted and observed lake area distribution. This model provides a slightly better fit to the distribution of small lower latitude lakes than the
uniform intermediate permeability model with enhanced runoff and recharge and evaporation suppression over Kraken Mare.

These model results indicate that subsurface flow is an essential component to the hydrological cycle on Titan. Subsurface flow directly accounts for approximately 1/3 of the influx into typical Titan lakes in our models. Moreover, subsurface flow plays a critical role in redistributing fluid across the surface over a range of spatial scales. Over short to intermediate length scales, subsurface flow acts as both a source and sink of fluid in order to produce the observed distribution of lakes. Infiltration and subsurface transport are necessary to prevent the formation and stability of a myriad of small lake perched at high elevation in every closed topographic depression. At the same time, subsurface flow provides a critical source of fluid to lakes, helping to stabilize them during periods of high evaporation so as to reduce the magnitude of seasonal changes. However, over long distances, the effect of subsurface flow is contrary to observations – long distance subsurface flow from temperate high latitudes to arid mid latitudes impedes lake formation in the arctic lake district. To balance these competing effects, either a uniform aquifer of intermediate permeability or a high permeability aquifer at the pole surrounded by an intermediate permeability aquifer is required to reproduce the sizes and latitudinal distribution of Titan’s lakes.

The models that best match the observed distribution invoke either a similar scaling law to partition precipitation between direct evaporation and recharge/runoff as that typically applied on Earth (uniform intermediate permeability model) or scale the recharge and runoff fraction by a factor of 3 (high permeability cap model). This suggests that the net effect of microscale processes in the soil layer on the fate of precipitation falling on the surface on Titan are of the same order of magnitude as on Earth. This inference is surprising given the important role of
biologically-derived organic material in Earth’s soils, and may suggest that atmospherically derived organics play a similar role in Titan’s soils.

Our work also reveals the critical role that some form of evaporation suppression over Kraken Mare plays in the stabilization of that sea, despite its spanning latitudinal climate zones ranging from temperate to hyper-arid. The necessity of a decreased evaporation potential over Kraken Mare in order to create a Kraken-sized sea suggests that Kraken Mare may be influenced by the effects of both ethane and/or dissolved organic solutes as suggested by previous studies (Tokano, 2009; Lorenz, 2014; Tan et al., 2015) and an evaporative cooling effect not accounted for at the dry, lower polar latitudes in the GCM, preventing appreciable evaporation over Kraken Mare. This problem is exacerbated by subsurface flow, which acts as a net sink of fluids flowing from Kraken to the surroundings in the lower latitude reaches of the sea.

Although this work has made significant advances in representing the full hydrological cycle of Titan, it nevertheless relies on a number of assumptions and simplifications. The true topography of Titan is not likely to be purely fractal in character. Erosion, deposition, and dissolution appear to have modified the landscape in a manner that would affect both the scale of overland flow and the relief surrounding lakes and seas. However, the approximate agreement between lengths of drainage basins in the model and on Titan and the immature fluvial state of much of the Titan landscape suggests that the fractal terrain is a reasonable simplification. Although preliminary models using diffused landscapes found similar results to those presented in this work, future work should include the effects of landscape evolution on Titan. Dissolution also appears to play a role in the formation of steep-sided lakes and may influence the distribution of small lakes. Given that the basin-scale models focused on the hydrology of the large seas, these small steep-sided depressions will have little influence on the model results for
the basin-scale models. The metric that the polar models were compared against (the latitudinal
distribution of lakes) is dependent on the balance of fluxes into and out of the model, which will
be controlled by the assumed recharge and runoff, and evaporation suppression. Thus, a lack of
explicit steep-sided depressions in the polar models will only have a slight influence on the
latitudinal distribution of lakes. Future work will investigate the influence of dissolution and
deviations from Darcy flow in the subsurface. The GCM included a sophisticated treatment of
the methane cycle, but did not include the effects of large seas on the climate. Future work
should include known lakes and seas in the GCM, as well as the effects of the surface
topography. Our models also assumed the liquid to be pure methane. Although the effect of
solutes on the viscosity is minor in comparison to the uncertainty in the permeability, solutes can
play an important role in the suppression of evaporation. Nevertheless, our simplified treatment
of evaporation suppression over Kraken Mare enabled us to quantify the magnitude of the
combined effects of the concentration of solutes in the sea and the impact of the sea on the local
climate. The conclusion that decreased evaporation over Kraken Mare remains robust even with
our simplified treatment of evaporation suppression.

In summary, while there is clear evidence for the surface and atmospheric components of
Titan’s hydrological cycle, there are no direct observations of the subsurface component. By
using models that integrate the atmospheric, surface, and subsurface components of Titan’s
hydrological cycle, and comparing the results to the observed sizes and spatial distribution of
lakes, we have made inferences regarding the nature of Titan’s subsurface and surface hydrology.
Subsurface flow plays a critical role in governing the sizes, distribution, and stability of lakes
and seas on Titan.
CHAPTER 3
RECONSTRUCTING THE PAST CLIMATE AT GALE CRATER, MARS FROM HYDROLOGICAL MODELING OF LATE-STAGE LAKES

Abstract. The growing evidence supporting the existence of a Gale Crater lake over a geologically significant period (100 to 10,000 years) raises further questions concerning Mars’s past hydrology and climate. The sedimentary deposits in Gale Crater may preserve one of the best records of the early Martian climate during the late Noachian and early Hesperian, and the transition from wetter conditions when fluvial valleys and lakes were active to drier conditions when most of the thick accumulations of sediments were forming. Remote sensing observations of fan deposits, inferred to be deltas, allow for estimates of the past paleo-lake levels in Gale Crater. Here we use numerical models of the surface and subsurface hydrology of Gale Crater and its surroundings, using estimates of past lake levels to put constraints on the past climate of Mars during the Early to Late Hesperian. Using Earth analog climates, I show that an aquifer permeability on the order of $1 \times 10^{-13}$ m$^2$ under a semi-arid to arid climate match the observed Gale lake stands. These results are comparable to climate for conditions similar to cold desert and steppe climates on Earth.

3.1 Introduction

The broad distribution of dendritic valley networks (Craddock et al., 1997; Craddock and Howard, 2003; Irwin et al., 2005; Howard et al., 2005) and abundant clay bearing units exposed in Noachian-aged terrain (e.g., Bibring et al., 2006; Mustard et al., 2008; Murchie et al., 2009) led to the idea that early Mars was warmer and wetter and likely experienced a hydrologic cycle that included precipitation-induced runoff (Hynek and Phillips, 2003). Morphology (Howard,
2007) and discharge estimates (Hoke et al., 2011) of Martian channels have interpreted semi-arid to arid past climate conditions during the Late Noachian and Early Hesperian, although other studies argue that the immaturity of the fluvial morphology is more consistent with hyper-arid environments on Earth (Stepinski and Stepinski, 2005; Irwin et al., 2011). Investigation of the interconnected surface hydrology using open-basin lakes and valley networks in the southern highlands (Matsubara et al., 2011; 2013) suggested a climate similar to the Pleistocene Great Basin region when conditions were wetter and colder during the last glacial maximum. Hydrated clay minerals (Mustard et al., 2008), fan deposits, and fluvial channels discharging into impact basins indicate that water existed on the surface as crater lakes (Forsythe and Blackwelder, 1998; Cabrol and Grin, 1999; Fassett and Head, 2008; Goudge et al., 2012; 2015) over timescales necessary to deposit and aqueously alter rocks in a lacustrine environment. This is followed by a transition in the late Noachian and early Hesperian to sedimentation, sulfate deposition, and potentially subsurface driven hydrology (Bibring et al., 2006). Evidence for fluvial activity into the Hesperian and possibly as late as the Amazonian (e.g., Mangold et al., 2008; Hynek et al., 2010) indicates that wetter climate conditions than those of the present day may have existed later than previously thought during short-lived episodes.

First identified as containing a potential closed-basin lake by Forsythe and Blackwelder (1998), Gale Crater is a Late Noachian to Early Hesperian crater located on the dichotomy boundary and is currently the location of the Mars Science Laboratory (MSL) on the Curiosity rover. Remote observations of Gale Crater noted layered deposits in a central crater mound (Aeolis Mons) that had been fluvially eroded, fan deposits on the crater floor, and fluvial dissection of the crater rim (Cabrol and Grin, 1999; Irwin et al., 2005; Anderson and Bell, 2010; Le Deit et al., 2013; Grotzinger et al., 2015; Palucis et al. 2014; 2016). Spectral analysis of the
Gale Crater mound deposit noted the presence of clays interlayered with sulfate-bearing deposits (Milliken et al., 2010; Thomson et al., 2011) suggesting intermittent times of dry and wet climates. Recent observations by MSL revealed further evidence for a past wet environment in Gale (e.g. Palucis et al., 2014; Grotzinger et al., 2014; Stack et al., 2014; Siebach et al., 2014, Palucis et al. 2016). Mudstone layers on the crater floor and deltaic deposits extending from the crater rim provided evidence for stable lakes over 100’s to 10,000’s of years based on assumed sedimentation rates similar to those on Earth (Grotzinger et al., 2014; 2015) possibly forming the basal unit of Aeolis Mons. Evidence for later stage hydrology, after mound formation, has been identified from fan deposits on the Gale Crater floor (Le Deit et al., 2013; Palucis et al., 2016).

Palucis et al. (2016) identified three distinct lake stands based on several Gilbert style deltaic deposits, similar to those found in lakes on Earth, and other immature fans breaking at roughly the same elevation within Gale Crater. These lake stands were estimated to fluctuate between -3280 m elevation (geoid referenced) based on the Pancake delta formation (followed by a desiccation period), -3980 m, and -3780 m consistent with the Farah Vallis delta. The large extent of these fan deposits suggest that these lakes persisted for 100,000 to 10,000 years depending on assumed discharge rates in the surrounding catchments and a rock to water ratio of 0.001. Open and closed-basin lakes, fluvial dissection, and aqueously altered minerals in the surrounding watersheds immediately outside of Gale Crater have also been identified (Cabrol and Grin, 1999; Fasset and Head, 2008; Goudge et al., 2012, 2015; Ehlmann and Buz, 2016). Although these studies provide evidence for a wetter climate in and around Gale Crater, the presence of lakes alone does not constrain the past climate as lake level is influence by the atmospheric, surface, and subsurface hydrology.
Although the existence of persistent liquid water ponded within Gale Crater both before and after the formation of Aeolis Mons is clearly demonstrated by the MSL observations, the specific nature of the climate cannot be constrained from the observations alone. Lake levels are a product of the complicated interplay of the surface and subsurface components of the hydrological cycle, which in turn are driven by the climate and the topography of a basin and its watershed. I use hydrological models of a Gale Crater to constrain a range of post-Aeolis Mons climates compatible with the lake stands inferred by Palucis et al. (2016). The models assume an average annual temperature above the freezing point of water, a surface hydrology similar to arid terrestrial regions, a regional contribution from a deep aquifer, and a range of climates based on scaled terrestrial climates. Evidence for aqueous alteration and hydrologic activity in craters surrounding Gale Crater are used as a further constraint of the past climate. I examine a range of climates representing different humidity regimes (e.g., hyper-arid, arid, semi-arid, sub-humid), as well as a range of assumptions in the hydrological model, in order to constrain the conditions on Mars around Gale Crater required to reproduce the observed post-mound lake levels.

3.2 Methodology

To constrain the climate and hydrogeology at the Gale Crater lake I used a hydrological model (Horvath et al., 2016), representing surface runoff in localized catchments, the subsurface flow in deep aquifers, and lakes. The lake stands of the modeled Gale Crater lake were then compared to the observed lake stands within Gale (Figure 3.1) for each model run. Models were run on MOLA geoid-referenced topography at a domain size of 620 km × 620 km and a model resolution of 3698 m. The size of the domain was chosen in order to prevent any boundary condition effects at Gale Crater and to capture several southern highlands craters outside of Gale Crater that have been identified as open-basin lakes (Cabrol and Grin, 1999; Fassett and Head,
This model was nested within a larger regional model extending from 60°E to 180°E, and 60°S to 60°N in order to capture any effects of longer-distance flow into and out of the model domain. This regional model neglected the effects of the spherical shape of the planet, but had only a second order effect on the results within the primary model grid.

Figure 3.1. MOLA topography of the Gale Crater region showing the inferred post-Aeolis Mons lake stands in Gale (Palucis et al., 2016) and identified open and closed-basin lakes outside of Gale (Cabrol and Grin, 1999; Fassett and Head, 2008; Goudge et al., 2012; 2015).

In order to investigate a range of climates, the model was forced at the surface using evaporation potential ($E_p$, the evaporation rate from a standing body of water) and precipitation ($P$) rates from Earth-based observations of analog climates provided by the North American Land Data Assimilation Systems Phase 2 (NLDAS-2; Xia et al., 2012a; Xia et al., 2012b). Earth climates were chosen to provide realistic daily and seasonal variations in precipitation and evaporation potential, analogous to a past Mars climate. Although multiple data sets were
acquired, I focus on a semi-arid, Great Plains climate from central Kansas and an arid, desert climate from the Gila River catchment to the southwest of Phoenix, AZ. Model results are most sensitive to the mean annual ratio of $E_p$ to $P$, referred to as the aridity index ($\phi$), and the annual precipitation. Thus, the semi-arid and arid climate datasets were scaled to investigate a range of aridity indices and annual precipitation. These climates are seasonally representative of an arid and a semi-arid climate, and allow for an investigation of a wide range of climate parameters while maintaining a consistent seasonal profile for each aridity range. For this study, I investigated an aridity index range between 1.5 and 33, representing aridity ranges between sub-humid and hyper-arid, while maintaining a constant annual precipitation of 250 mm/yr. At aridity indices above 10, this annual precipitation is unrealistically high for climates on Earth. For the sake of consistency, I maintain this constant annual precipitation to investigate the lake behavior for different aquifer permeability values and aridity. I also investigated a range of annual precipitation from 50 to 1500 mm/yr, encompassing climates ranging from a hyper-arid desert climate similar to the Arabian and Namib Deserts (~100 mm/yr) to a semi-arid steppe climate, similar to central Kansas (~600 mm/yr). I emphasize that my model does not explicitly take temperature into account, depending only on the rates of evaporation and precipitation. The model does require mean annual temperatures above freezing in order to allow for a vertically integrated hydrological cycle in which infiltration is not impeded by a continuous permafrost layer.

Surface boundary conditions to the groundwater and surface runoff models were determined from the climate data using an empirical method from terrestrial hydrology known as the Budyko relationship. Budyko-type estimates of evaporation from watersheds approximate the micro-scale processes at the surface-atmosphere interface using empirical discharge data from
terrestrial basins. This method derives a functional relationship for the actual evaporation within
a given basin dependent on the aridity index and a shape parameter ($\omega$; Zhang et al., 2004).
Higher shape parameter values result in higher actual evaporation from a catchment, allowing
less water to reach the surface and subsurface hydrologic system. Shape parameters for
individual catchments on Earth are generally determined using known discharge, and
precipitation rates. In general, higher shape parameters ($\omega > 2$) are attributed to grasslands and
forested catchments and lower shape parameters ($\omega < 1.7$) generally correspond to catchments
with properties not favorable to evaporation including high intensity seasonal precipitation, soil
conditions with low storage capacities, slope, and regions that lack vegetation (Zhang et al.,
2004). This study found an average shape parameter for grassland catchments of 2.5, consistent
with the shape of the original Budyko-relationship (Budyko, 1974) although subsequent studies
including groundwater flow for the continental United States found a median shape parameter of
1.8 with a range between 1.5 and 2 (Condon and Maxwell, 2017). A specific shape parameter,
though, cannot be attributed to specific climates due to the numerous properties that influence
the shape parameter. Thus for this study, we investigated a wide range of shape parameters from
1.3 to 2.6, although only results between 1.3 and 2 are presented in this work, which we believe
best represents surface hydrology conditions on Mars consistent with a lack of vegetation, high
topographic relief, and a lack of soil organics. Water that does not evaporate back into the
atmosphere is partitioned into either deep aquifer recharge or surface runoff to lakes. While
multiple processes may control runoff and channelized flow, the large time steps associated with
numerical subsurface modeling in comparison with the timescales for dynamic surface runoff
necessitate the use of a simple analytic approximation to represent runoff. For this study, a linear
reservoir approximation is used to determine surface runoff based on the assumption that the
storage in a catchment is linearly related to the runoff in the catchment (e.g., Cunge, 1969; Overton, 1970; Dooge, 1973), and is approximated using basic parameters for a given catchment. Water that recharges the aquifer will diffuse in the subsurface, which is modeled using a finite-difference approximation to the groundwater flow equation, dependent on the permeability distribution, aquifer depth, and intrinsic properties of water. Lake formation was allowed to occur where the water table intersected the surface. Ponded surface water was then run through a lake diffusion scheme in order to bring each lake surface to an approximately constant hydraulic head. This lake treatment solved the groundwater flow equation at lake pixels over multiple iterations for each iteration of the subsurface model, with the porosity of the lake surface set to 100% and a high permeability value of $10^{-10}$ m$^2$, in order to allow the lake surface to approach an equipotential.

For the subsurface model, we assumed a laterally homogenous aquifer in which the vertically averaged permeability from the surface to 10 km depends on depth based on the megaregolith aquifer model (herein referred to as the nominal aquifer model) of Hanna and Phillips (2005). This model assumed a brecciated megaregolith overlying a fracture basement. The Gravity Recovery And Interior Laboratory (GRAIL) data from the moon found that the highly impacted lunar surface is more porous than previously though (Besserer et al., 2013), which may support a higher permeability for the heavily impacted southern highlands on Mars. For the purpose of this work, we investigate a range of vertically averaged permeability from the surface to 10 km between $3\times10^{-14}$ m$^2$ to $3\times10^{-11}$ m$^2$ an order of magnitude lower and two orders of magnitude higher than the nominal aquifer model ($3\times10^{-13}$ m$^2$). Each aquifer permeability model follows the same decrease with depth as the nominal aquifer model, with a $100\times$ decrease in vertically averaged permeability from the surface to a depth of 5 km.
3.3 Lake stand dependence on climate and permeability

The area and surface elevation of a lake in Gale Crater is primarily dependent on the aridity index \( \frac{E_p}{P} \) (Figure 3.2). For the nominal permeability model, sub-humid conditions at an aridity index below 2 will over-predict (Figure 3.2a), while arid conditions at an aridity index above 5 will under-predict (Figure 3.2c) the highest observed lake stand in Gale Crater. Depending on the amount of water assumed to participate in the surface and subsurface hydrology (as determined by the shape parameter, \( \omega \)) an aridity index range between 3 and 4 matches the highest lake stand at -3280 m elevation for the nominal permeability model at an annual precipitation of 250 mm/yr (Figure 3.2b; 3.3a). An aridity index range between 4 and 6, at the transition between a semi-arid to arid climate, fits the lowest inferred lake stand at -3980 m elevation and an aridity index range between 3.5 and 5 fits the middle lake stand at -3780 m elevation, proposed as the final major lake stand in Gale. The aridity index range determined here (between 6 and 3) corresponds to a total recharge and runoff percentage between 21% and 31% of the annual precipitation. At an annual precipitation of 250 mm/yr, the low end of this range of aridity indices (3) is comparable to a temperate semi-arid climate, similar to the Eurasian steppe and western Kansas; whereas the high end of the range (6) is comparable to a mid-latitude, cold desert climates, similar to the Great Basin desert in Utah and the eastern slope of the Andes mountain range in Patagonia. The nominal permeability model predicts lake formation in the majority of southern highland craters in the proximity of Gale Crater as well as ponding in topographic lows in the northern lowlands and in Sharp crater to the west of Gale, found to have fan deposits and a similar mineralogy as Gale Crater (Ehlmann and Buz, 2015). The restriction of subsurface flow in these models causes a steep hydraulic head gradient and a near-surface water table, forcing water to the surface in local topographic lows. Although
widespread aqueous mineralization is not observed in the northern lowlands, fluvial systems at the dichotomy boundary (Kite et al., 2015) that would have channeled much of the fluid that forms the northern lowland lakes away. Present-day topography includes partially eroded deposits, which will assist in ponding in the northern lowlands thus much of the ponding observed in this region may not have actually occurred.

Although the aridity index has the strongest influence on the Gale Lake elevation, the annual precipitation will also influence the distribution of lakes inside and outside of Gale. Decreasing the annual precipitation for a given aridity index (and thus decreasing the evaporation potential) results in a decrease in aquifer recharge, a deeper aquifer, and a reduction in lakes outside of Gale, and a larger stable lake in Gale Crater. At the lowest annual precipitation investigated here (94 mm/yr), an aridity index in the arid climate regime (6) matches the lowest lake stand in Gale, although a semi-arid climate (4) is still required to match the highest lake stand, even at this low precipitation (Figure 3.3b). In contrast, an annual precipitation higher than the 250 mm/yr previously investigated predicts a shift towards lower aridity indices, and a semi-arid climate range is still predicted, similar to a central Kansas climate at an annual precipitation of 600 mm/yr. However, we find that the absolute precipitation rate has a much weaker influence on the outcome of the models than the aridity index.

The aquifer permeability will also influence Gale Crater lake elevation and distribution of lakes outside of Gale (Figure 3.4). As permeability is increased, subsurface flow becomes a prominent source to the Gale Crater lake, allowing a given lake stand to be achieved for more arid climates. However, this also has the effect of reducing the lake area outside of Gale Crater, similar to the effects of lowering the annual precipitation (Figure 3.4b; 3.4c). The high
Figure 3.2. Lakes (shown in black) and hydraulic head maps (contour) overlain on Gale Crater topography for different aridity indices of a) 1.5, b) 3.5, and c) 9. All models shown use an annual precipitation 300 mm/yr, a $k=3\times10^{-13}$ m$^2$ and a shape parameter of 1.6. Models a) and c) over and under-predict the highest observed lake stand respectively (shown as the white outline), while b) matches the highest observed lake stand and predicts ponding in several of the identified open-basin lakes outside of Gale.
Figure 3.3. The dependence of the Gale Crater lake elevation on the aridity index is shown for a) several shape parameters at an annual precipitation of 250 mm/yr and an aquifer permeability of $3\times10^{-13}$ m$^2$, b) different annual precipitation values for a shape parameter of 1.6 and a permeability of $3\times10^{-13}$ m$^2$, and c) different aquifer permeabilities at an annual precipitation of 250 mm/yr and a shape parameter of 1.6. The horizontal dashed lines are the lake stands inferred by Palucis et al. (2016) and correspond to the colored contours in Figure 3.1.
permeability and lack of lakes outside of Gale Crater in these models enhances subsurface flow to Gale by tapping an extensive recharge zone, driving long distance subsurface flow resulting in a stable Gale Crater lake for drier climates. For an aquifer model with higher permeability than the nominal model, arid climates fit the range of observed lake stand in Gale Crater. An aridity index of 12 with an annual precipitation of 94 mm/yr, comparable to cold desert climates like the Taklamakan desert in western China and the Gobi desert in Mongolia, fits the lowest Gale lake stand for the higher permeability $10 \times$ the nominal model (Figure 3.3c). An aridity index of 5, at the transition between semi-arid and arid climates, matches the highest lake stand, at an annual precipitation of 250 mm/yr. Likewise, a permeability $100 \times$ the nominal model matches the lowest lake stand at an aridity index between 15 and 20 and the highest lake stand at an aridity index of 9, well into the arid climate range (Figure 3.3c). However, these high permeabilities fail to meet the constraint imposed by geological evidence for lakes within a number of craters outside of Gale. In contrast, low permeability models predict a shallower water table and more lakes outside of Gale. Although it is difficult to argue against the presence of numerous small lakes as predicted by these models, there is no geological evidence supporting them. While permeability has a larger impact on the climates required to support the Gale Crater lake, higher and lower permeability models are difficult to reconcile with evidence for lakes outside of Gale.

### 3.4 Discussion

Results favor the nominal aquifer permeability, which predicts a semi-arid climate for the post-Aeolis Mons lakes, possibly reaching arid climates for the lower lake stands depending on the assumed annual precipitation and shape parameter. A semi-arid climate is largely consistent with discharge estimates from Martian channels (Hoke et al., 2011). The nominal aquifer model predicts lakes in southern highland craters that roughly match the observed open-basin lakes
Figure 3.4. Lakes (shown in black) and hydraulic head maps (contour) overlain on Gale Crater topography for different aridity indices and permeabilities of a) 4 and $3 \times 10^{-13}$ m$^2$, b) 5 and $3 \times 10^{-12}$ m$^2$, and c) 9 and $3 \times 10^{-11}$ m$^2$. All models shown use an annual precipitation 300 mm/yr and a shape parameter of 1.6. All models match the highest observed lake stand (shown in the white outline) while the lake distribution outside of Gale differ.
(Cabrol and Grin, 1999; Fassett and Head, 2008) for annual precipitation between 100 and 600 mm/yr and an aridity index between 3 and 6, corresponding to climates similar to Eurasian steppe or western Kansas for the high lake levels, and the Great Basin desert in the United States or eastern slope of the Andes mountain range in Patagonia for the low lake levels. Although an aquifer permeability on the low end of the permeability range investigated here is favored, subsurface flow is a significant contribution to the inflow of water to the Gale crater lake, accounting for ~39-47% of the total influx for the highest lake stand depending on the amount of recharge and runoff assumed. A subsurface contribution near 50% to a Gale crater lake has implications for solute transport to Gale and the surrounding regions where chloride deposits have been observed (Ehlmann and Buz, 2016). Sulfate deposits in the central mound (Milliken et al., 2010) may have been formed in a playa environment, which would require a substantial subsurface flow contribution and solute concentration to form the thick sulfate layer observed, in agreement with the large subsurface contribution to Gale.

While models do not explicitly take into account, and the relationship between temperature, precipitation, and evaporation potential are likely not the same for early Mars and present Earth, in general an increase in temperature or a decrease in water to the system is required to shift the climate towards more arid conditions. For aquifer permeability values greater than the nominal aquifer model, greater aridity indices are required to fit the observed lake stands and higher annual precipitation is needed to produce crater lakes outside of Gale. Given that climate models currently have difficulties raising early Mars temperatures above the freezing point of water (e.g., Ramirez et al., 2014; Wordsworth et al., 2015) the warmer conditions implied by the higher precipitation rate and higher aridity index is unfavorable for an early to late Hesperian Mars climate.
The relatively narrow range of aridity indices that match the observed lake stands for the nominal model has interesting implications for climate change between lake stands. Although drier conditions are required to match the lowest observed lake stand compared to the highest, a drastic change in the climate is not required. One scenario, following the post-mound lake stands chronology of Palucis et al. (2016), forms a lake at the highest lake stand at relatively wet semi-arid conditions, similar to a central Kansas climate. Limited water availability result in the drier, semi-arid climates, similar to montane deserts or sub-polar regions, that form the lakes at the lowest and middle lake stands.

3.5 Conclusions

This work provides constraints on the past early to late Hesperian climate at Gale Crater using a hydrologic model and indicators of paleo-lake levels. These climatic inferences cannot be made based on the lake levels alone, since lake levels are controlled by the complicated interplay of the surface and subsurface hydrology, and by the effects of the unique topography surrounding this crater situated on the dichotomy boundary. Inferred climates in the semi-arid to arid range at this time in the Hesperian are consistent with previous estimates of the climate on Mars during the Noachian based on geomorphology and the properties of the fluvial systems. This convergence of multiple approaches on a similar climate lends strength to the conclusions. Although much work remains to be done to work out the details of Mars’ climate evolution, and debate continues between warm-wet and cold-icy scenarios, this work shows that these late lake stands in Gale Crater are consistent with semi-arid climates with active hydrological cycles that require mean annual temperatures above freezing. With continued observations of the extent and elevation of lake bed deposits from MSL as it climbs Aeolis Mons, I can further constrain the past Martian climate, piecing together the climate history and hydrogeology of Mars.
CHAPTER 4

IMPLICATIONS FOR THE EARLY CLIMATE OF MARS FROM HYDROLOGICAL MODELING OF LAKES AND LAKE DEPOSITION AT GALE CRATER, MARS PRIOR TO THE FORMATION OF AEOLIS MONS

Abstract. The surface of Mars holds evidence for conditions much wetter than the current cold and dry surface observed today. Gale Craters central sedimentary mound, formally known as Aeolis Mons, and other related sedimentary deposits on Mars, may provide the best observational evidence for climatic conditions of past Mars and the transition from wet to dry conditions. Observations by the Mars Science Laboratory (MSL) on the Curiosity rover during its traverse up Aeolis Mons provide ground-truth evidence of lacustrine deposits in the base of Aeolis Mons, and a transition in depositional environments recorded at higher levels. In this work I focus on deposition in a lacustrine environment in Gale crater to form the basal layers of Aeolis Mons. I investigate the lake hydrology of Gale crater prior to the formation of Aeolis Mons, and the influence climate and topography on the hydrology. Semi-arid climates predict deep large lakes in a pre-Aeolis Mons Gale that match the elevation of observed lake deposits, embaying the central peak and also forming lakes in craters known to have hosted lakes outside of Gale. I also find that deposition in a lacustrine environment up to the level of the observed clays in the mound has little effect on the formation of lakes in Gale. Thus, in order to account for the inferred change in depositional environments from lacustrine to evaporitic settings observed in the sediment package, a change in climate is necessary.
4.1 Introduction

Gale Crater is a 154 km diameter crater located at 5.4°S, 137.8°E on the boundary between the northern lowlands and southern highlands. The degradation of Gale (Forsberg-Taylor et al., 2004; Irwin et al., 2011), superposed cratering on Gale ejecta (Thomson et al., 2011), and crater-size frequency distribution on the crater floor (Thomson et al., 2011) puts the formation of Gale between the Late Noachian to Early Hesperian (3.8 to 3.5 Ga). Gale crater has been of particular interest due to the 5.5 km-tall sedimentary mound, known as Aeolis Mons, which may preserve one of the best records of the early Martian climate during the transition from wetter conditions when fluvial valleys and lakes were active to drier conditions when thick accumulations of sediments were forming (Bibring et al., 2006; Murchie et al., 2009). The centrally located mound consists of fine grained layers of sedimentary material with two distinct mineralogical units (Milliken et al., 2010; Thomson et al., 2011). Aqueously formed clays and sulfates make up the lower portion of the mound, extending up to ~3400 m elevation, with estimated dips <2°. A systematic transition from hydrous to anhydrous sediments in the mound consists of a 400 m basal clay member (Grotzinger et al., 2015), a 300 m sulfate layer with weak clay signatures, and a 400 m sulfate layer although weak sulfate signatures are observed at higher levels of the mound (Milliken et al., 2010; Thomson et al., 2011). In contrast, the upper mound unit overlaying an erosional discontinuity with the lower unit, is primarily aeolian influenced and shows no spectral evidence of hydrated minerals. In the lower mound, there is evidence for a change in the aqueous environment, from a more neutral-pH lacustrine environment indicated by clay deposits to a more acidic environment indicated by intermixed with sulfate and aeolian deposits. The gradual transition from more clay rich layers at the base of the mound to sulfate-bearing units within the bulk of the mound (Milliken et al., 2010) is consistent with the layering
observed in Meridiani Planum (Glotch and Rogers, 2007) and Arabia Terra (Zabrusky et al., 2012) suggesting a similar formation mechanism for layered mound deposits (Andrews-Hanna et al., 2007; Andrews-Hanna and Lewis, 2011). Of the numerous formation mechanisms that have been proposed for mound formation, aeolian deposition and erosion (Kite et al., 2013), lacustrine deposition and subsequent erosion of crater fill (Grotzinger et al., 2015), and evaporite playa deposits (Andrews-Hanna et al., 2007) are the most widely accepted. The formation of the sulfate layers within Aeolis Mons due to aqueous cementation of aeolian sediments by a fluctuating water table would suggest that the crater was at one time filled with sediments, which were later eroded into the present-day central mound. Alternatively, other studies suggest an aeolian origin for the mound in its current geometry, pointing to the bedding dips and the difficulty in removing crater fill material (Kite et al., 2013).

Thus far the Curiosity rover, carrying the Mars Science Laboratory (MSL), has driven over two distinct groups; the Bradbury group, which composes the current floor unit of Gale Crater, and the Murray mudstone group, which was interpreted to be the basal unit of Aeolis Mons (Grotzinger et al., 2015). The Bradbury group is composed of interlayered conglomerates (Williams et al., 2013) and fine grained sandstone sediments with dips away from the crater rim, which may represent deltaic deposits. Mudstone layers at the base of Aeolis Mons, composing the Murray group are interfingered with the Bradbury group and are postulated to be the deposits of a past Gale Crater lake (Grotzinger et al., 2015). Based on a thickening of coarser grained deltaic deposits, a greater number of finer grained mudstone layers along the Curiosity traverse, and dips toward Aeolis Mons, this study suggests that the transition from the Bradbury group to Murray group represents a change in the depositional environment of a past Gale Crater lake from deltaic deposits at the margins to finer grain deposits in a distal, quiescent portion of the
lake. Unlike later lakes revealed by orbital observations of deltas along the crater rim and central mound that post-date the erosion of the central deposit to its current mound shape (Palucis et al., 2016), the lake revealed by the Bradbury formation and Murray group formed shortly after the crater was formed and pre-date the sedimentary infilling of the crater.

This work focuses on the nature of the hydrology and climate required to sustain a lake within Gale Crater as indicated by the prominent mudstone layer observed by Curiosity. Previous hydrological modeling work (Andrews-Hanna et al. 2007; Andrews-Hanna and Lewis, 2011) was focused on the formation of sulfate-rich sedimentary deposits in Meridiani Planum and Arabia Terra, similar to those making up the bulk of Mount Sharp. That work identified preferential sites of groundwater upwelling and evaporite deposit thickness based on a simplified hydrologic cycle. Gale crater was shown to be a unique site for hydrological activity, as it is located on the dichotomy boundary between the highlands and lowlands and has a large recharge zone that will contribute to enhanced groundwater flow (Andrews-Hanna et al., 2012). That work was able to explain the formation of the observed sulfate deposits, but could not capture aspects of the surface hydrology such as runoff and ponding necessary to explain the lacustrine deposits at the base of Aeolis Mons. Localized modeling of the post-Aeolis Mons hydrology at Gale including overland flow and ponding (Chapter 3) found that climate, specifically the aridity index, and permeability had the largest influence on the areal extent of a lake in Gale.

Here, I use a hydrological model localized at Gale crater to investigate the climate and hydrological conditions necessary for the deposition of older lake sediments making up the base of Aeolis Mons. MSL observations require lakes extending at least up to the top of the Murray mudstone at -4200 m. The fact that this deposit interfingers with the Bradbury formation, interpreted as delta deposits, indicates that the paleo-lake surface was close to this level, so I take
this as the target lake depth for our models. Sedimentary deposition in Gale can influence the formation of lakes and depositional environment in Gale crater. Although a prominent change in the nature of the deposits (clay to sulfates) and the inferred depositional environment (lacustrine to playa) is observed within Aeolis Mons, such a change could possibly result from the changing hydrology in response to the infilling of the crater rather than a climate change. To simulate the different depositional time periods in Gale Crater, I first reconstruct Gale crater’s topography prior to the formation of Aeolis Mons, and then use a simple fill model to represent added lake sediments to investigate the influence of crater infill on lake formation for a range of past climates. I then investigate a range of climate in order to constrain which are capable of reproducing the lake levels inferred from the observed lacustrine deposits.

4.2 Methodology

In order to investigate the influence of crater infill and climate on the formation of a Gale Crater lake prior to the deposition of Aeolis Mons, we use a numerical hydrological model developed by Horvath et al. (2016) that incorporates the atmospheric, surface, and subsurface components of the hydrological cycle. Models were run on MOLA geoid-referenced topography at a domain size of 620 km × 620 km and a model resolution of 3698 m (Figure 4.1a), using topographic models of the pre-Aeolis Mons Gale and subsequent crater infill topography. This model was nested in a larger regional model extending from 60°E to 180°E, and 60°S to 60°N in order to capture any effects of longer-distance flow into and out of the model domain. Earth-based evaporation potential and precipitation data for arid and semi-arid climates were used to force the models, which in turn control aquifer recharge and surface runoff. Lakes were allowed to develop naturally based on the resultant surface and subsurface flow as determined by a numerical model.
4.2.1 Pre-Aeolis Mons topographic model and crater infill

Fluid flow in an unconfined aquifer is largely controlled by the topography. In order to model the pre-Aeolis Mons hydrology at Gale Crater, the central mound was removed and a central peak from a similar sized complex crater was added. Although the formation of a crater on the pre-existing topographic dichotomy is complex, we assume that the slope of the crater floor between the southern and northern crater rim are outcomes of the crater’s formation on the dichotomy boundary. Thus, besides the removal of the central crater mound and addition of a central peak, the topography of Gale Crater is largely left unaltered. Removal of the central sediment mound was done using a mask over the central mound based on the contact between the floor unit and the basal unit of the mound (Grotzinger et al., 2015). An inverse distance weighted interpolation algorithm was then used to create a smooth surface in the crater interior, removing both the mound and central peak. In order to include a realistic central peak in a pre-Aeolis Mons Gale, a central peak crater (located at 13°S, 204°E) of similar size was selected and a matching mask was constructed. To integrate the central peak in this crater with the topography of Gales floor, an inverse distance weighted interpolation at the edge of the central peak in the second crater was performed, and the elevation of the central peak above this surface was determined and then added to the smoothed floor of Gale crater. The central peak was centered at the proposed location of the mound-covered central peak in Gale (Grotzinger et al., 2015), located at the southern edge of the mound. The resulting topography is shown in Figure 4.1b. Although the central peak does not reach to the level of the proposed central peak in Gale, the highest elevation of the peak is above the lowest level of the crater rim and thus will not affect the infill models investigated in this study.
The presence of potential lake deposits at the base of the sediment mound and deltaic deposits suggests that the construction of the basal unit of the central mound likely occurred in a standing body of water. Sedimentary deposition after the formation of the crater would affect the subsequent hydrology and lake formation, and thus I also must consider partially filled craters in our analysis. Crater infill was modeled in this study using a simple landscape diffusion model. This approach used a differential eight-point method with a diffusivity constant set at 0.001 and only allowed infill to occur at predefined levels in Gale Crater. While multiple landscape diffusion models were explored, the landscape diffusion model used in this work only allowed diffusion below a fixed point set at different elevations on the crater wall (Figure 4.2). While deposition is inherently more complicated than assumed here, these simple models allow us to explore the different conditions of crater infill and investigate their effects in the lake level. It should also be noted that the depth of the infill explored here is beyond the level of the aqueously altered minerals in order to investigate the effect that the full range of possible sedimentary fill has on lake stability. The greater fill thicknesses are important to test whether the observed
transition from clays to sulfate within the sediments (and the inferred transition from lacustrine
to playa environments) could be a result of the infilling of the crater rather than a result of a
change in climate.

![Figure 4.2. A topographic profiles through pre-Aeolis Mons Gale Crater and subsequent infill based on the diffusive model with layered infill.](image)

4.2.2 Climate

The model was forced at the surface using evaporation potential ($E_p$) and precipitation ($P$) rates from Earth-based observations of analog climates provided by the North American Land Data Assimilation Systems Phase 2 (NLDAS-2; Xia et al., 2012a; Xia et al., 2012b). Although multiple data sets were acquired, we focus on a semi-arid, Great Plains climate from central Kansas and an arid, desert climate from the Gila River catchment to the southwest of Phoenix, AZ. Model results are most sensitive to the mean annual ratio of $E_p$ and $P$, referred to as the aridity index ($\phi$), and the annual precipitation. Thus, the semi-arid and arid climate datasets were scaled to investigate a range of aridity indices and annual precipitation rates. These climates are seasonally representative of an arid and a semi-arid climate, and allow for an investigation of a wide range of climate parameters while maintaining a realistic seasonal profile.
for each aridity range. For this study, we investigated an aridity index range between 3 and 24, representing aridity ranges between semi-arid and hyper-arid, while varying precipitation to represent Earth climates for these aridity indices. The scaled Kansas climate dataset was used for aridity indices equal to and less than 5 and the scaled Arizona climate dataset was used for aridity indices greater than 5.

Of the total precipitation falling on the surface, only a fraction contributes to runoff and recharge, while the rest evaporates back into the atmosphere. Surface boundary conditions for the groundwater and surface runoff models were determined using an empirical method from terrestrial hydrology known as the Budyko relationship (Budyko, 1974). Budyko-type estimates of evapotranspiration (i.e., evaporation from the bare surface and transpiration from vegetation) from watersheds approximate the net effects of micro-scale processes at the surface-atmosphere interface using empirical discharge data from terrestrial basins. This method derives a functional relationship for the actual evapotranspiration within a given basin dependent on the aridity index. The most widely used form of the Budyko relationship fits data from watersheds with the functional form:

\[ E_a(\phi) = P \left[ 1 + \phi - (1 + \phi^\omega)^{1/\omega} \right] \]  

(1)

where \( \omega \) is an empirically derived parameter, known as the shape parameter, dependent on the topography, soil properties, and vegetation (Zhang et al., 2004). Water that does not evaporate back into the atmosphere is partitioned into either deep aquifer recharge or surface runoff to lakes. In this work we use an arbitrary 50\% of the excess precipitation for runoff and recharge. At the large temporal and spatial scales used in this work Horvath et al. (2016) showed that the amount that recharges or runs off has little effect on the stability of lakes.
4.2.3 Aquifer model and groundwater flow in a porous medium

We assume a laterally homogenous aquifer in which the vertically averaged permeability from the surface to 10 km depends on depth based on the megaregolith aquifer model (herein referred to as the nominal aquifer model) of Hanna and Phillips (2005). This model assumed the permeability is dominated by the effects of fractures, which close with depth as revealed in Earth-based studies. This model has a vertically averaged permeability of $3 \times 10^{-13}$ m$^2$ from the surface down to 10 km, with a 100× decrease in the vertically averaged permeability down to 5 km depth. The vertically averaged permeability controlling flow at a given location depends on the depth of the water table beneath the surface. The porosity is assumed to follow an exponential relationship:

$$n(h) = n_0 e^{-[(z-h)/d_0]}$$

with a surface porosity ($n_0$) of 0.2 and a scale height ($d_0$) of 2.8 km (Binder and Lange, 1980; Clifford and Parker, 2001; Bahr et al., 2001).

Subsurface flow in a deep unconfined aquifer was modeled using a finite-difference approximation to the Boussinesq approximation of the groundwater flow equation:

$$\frac{\partial}{\partial x} \left( K_x b \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left( K_y b \frac{\partial h}{\partial y} \right) = n \frac{\partial h}{\partial t}$$

where hydraulic conductivity ($K$) is directly related to the permeability ($k$), gravity ($g$), and the intrinsic properties of water (density, $\rho$ and viscosity, $\mu$),

$$K = \frac{k \rho g}{\mu}$$

The active aquifer thickness, $b$, varies in space and time as a function of the hydraulic head $h$ and the topography $z$:

$$b(x, y, t) = h(x, y, t) - z(x, y) + d$$
in which \( d \) is the total thickness of the aquifer (assumed to be a constant value of 10 km).

### 4.2.4 Surface water

The large time steps associated with numerical subsurface modeling in comparison with the timescales for dynamic surface runoff necessitate the use of a simple analytic approximation to represent runoff. For this study, a linear reservoir approximation is used to determine surface runoff based on the assumption that the storage in a catchment is linearly related to the runoff in the catchment (Cunge, 1969; Overton, 1970; Dooge, 1973). We assume that the storage in a catchment is controlled by the excess precipitation that reaches the surface (as determined by the Budyko-type relationship discussed in Section 4.2.2), the discharge at the catchment outlet, and the time it takes a parcel of water to traverse the catchment. The geometry of each catchment was determined using an eight-point routing method (Wang and Hjelmfelt, 1998), which routes water through the catchment based on the steepest downslope values determined using the eight neighboring elevation points from MOLA topography.

Liquid that ponds on the surface in these models was run through a lake diffusion scheme in order to bring each lake surface to an approximately constant hydraulic head. This lake treatment solved Eq. (3) at lake pixels over multiple iterations for each iteration of the subsurface model, with the porosity of the lake surface set to 100% and a high permeability value of \( 10^{-10} \) m\(^2\). The lake diffusion was only applied to pixels where the water table was above the surface, although liquid was allowed to diffuse over the surface as needed.
4.3 Results

4.3.1 Pre-Aeolis Mons Climate

Here we investigate the conditions required for the pre-mound lakes. Desert climates, comparable to Arizona and the Arabian desert ($\phi = 9$, $P = 300$ mm/yr and $\phi = 24$, $P = 100$ mm/yr respectively), only form lakes in the lowest regions within Gale (Figure 4.3a; 4.3b). In contrast, climates that approach semi-arid conditions ($\phi = 6$, $P = 300$ mm/yr) form a continuous lake encompassing the northern half of Gale Crater and embays the northern portion of the central peak (Figure 4.3c). Unlike the desert climate conditions, a semi-arid climate predicts lakes in craters outside of Gale, many of which have been identified as open or closed-basin lakes (Fassett and Head, 2008). Similar lake conditions in Gale and the surrounding regions exist for a central Kansas climate ($\phi = 3.5$, $P = 600$ mm/yr) with slightly more saturation present in southern highland craters and the northern lowlands (Figure 4.3d). For an aridity index below 6, lake elevation is at or above -4200 m for the pre-Aeolis Mons topography, which coincides with the top of the Murray mudstone layer (Figure 4.4). Although dry climate can form small lakes in the deepest depressions within Gale, climates with aridity indices above 6 fail to reach the level of the Murray mudstone layers above -4200 m.

4.3.2 Influence of crater infill

Models of the hydrology for the reconstructed pre-mound Gale topography represent the earliest stages of lakes in the crater, but the lakes themselves would have contributed to sedimentary infilling of the crater, as revealed by the mudstones on the crater floor. The amount of infill and infill geometry can have a substantial influence on the lake level in Gale Crater. For
Figure 4.3. Lake maps (shown in black) overlain by hydraulic head maps (contour) on modeled pre-Aeolis Mons topography. Decreasing aridity indices from a) 24, b) 9, c) 6, to d) 3.5 are shown for the nominal permeability model.

the assumed diffusive fill geometry, the predicted lake elevation largely parallels the crater fill level for a constant aridity index, while lake area remains relatively constant as the level of crater infill increases (Figure 4.4a). The hydrological balance of the lake is controlled by the total evaporative loss from the lake surface, which scales with the area of the lake. Thus as the infill level rises lake level correspondingly rises to maintain a constant lake area. Even under dry
conditions, at an aridity index of 24, a relatively small and shallow lake in the northern portion of Gale Crater forms with a depth of ~50 m as the lake infill level increases. Lake area though is <1000 km² (Figure 4.4b) and does not fully encompass the crater or reach the contact between the crater floor unit and the basal unit of the mound. For an aridity index of 3.5 and an annual precipitation of 600 mm/yr, similar to a central Kansas climate, deep lakes that embay the central peak persist up to infill levels at the northern crater rim and maximum lake depths of ~1 km are predicted for the wettest climates.

4.4 Discussion

Reconstructing the pre-Aeolis Mons topography assuming the same crater floor slope observed at present, requires climates in the wetter arid and semi-arid climate range ($\phi \leq 6$). Although climates with an aridity index less than 6 predict lake levels above the observed Murray mudstone and sulfate contact, wetter climates cannot be ruled out. This puts the
formation of the Murray mudstones in a semi-arid climate, comparable to central Kansas or the eastern Great Basin desert in Utah, which is largely consistent with results from previous studies (e.g., Hoke et al., 2011; Horvath and Andrews-Hanna, 2017).

For an infilling scenario that assumes diffusive topography, lake depth and lake area are only slightly influenced by the increased infill in Gale Crater, while the lake elevation closely follows the level of infill. Wetter climates predict deep lakes in Gale Crater while drier climates predict shallow lakes (<100 m) for aridity index >9, similar to deserts on Earth. At all climates, lake depths remain fairly consistent as a function of fill depth, suggesting that the construction of the lower portions of the mound through lake deposition cannot account for the change from wet to dry conditions observed in the sediment packages. Due to Gale crater's unique location at the dichotomy boundary, making it exceptionally deep relative to the southern highlands, shallow lakes can form even under hyper-arid conditions in Gale. Thus, if lakes could be ruled out during the deposition of the sulfates and ash in the upper levels of the mound, this would necessitate either conditions at the surface that are no longer in communication with the subsurface aquifers or a parched climate where seasonal rainfall is nonexistent. More likely, small scattered ephemeral lakes could have existed during the deposition of the sulfate deposits, similar to the interpretations of standing or flowing water at Meridiani (Grotzinger et al., 2005).

4.5. Conclusions

This work provides constraints on the past late Noachian climate at Gale crater prior to the formation of Aeolis Mons using. Using a hydrological model and the lake sediments observed in the lowest basal layer of the mound, an upper limit on the aridity, at the transition between arid to semi-arid, is found. This is in agreement with past morphological studies (Stepenski and Stepinski, 2005; Howard, 2007) and discharge estimates (Hoke et al., 2011) of
southern highland fluvial networks. These climate conditions also predict lake formation in craters outside of Gale in general agreement with the distribution of closed- and open-basins in the southern highlands surrounding Gale (Cabrol and Grin, 1999; Fassett and Head, 2008). Lakes can form under hyper-arid conditions but the areal extents of these lakes in Gale are small and lake depths are shallow. Models accounting for the deposition of sediment in Gale predict consistent lake depths and areas, with the lake elevation roughly following the infill of the crater. Under all conditions investigated lakes still form as the crater infills due to the unique location of Gale crater on the dichotomy boundary. This suggests that the transition from hydrous layers in the lower mound to dry conditions recorded in the bulk of the mound requires a drastic change in the climate at Gale. On the other hand, if lakes were present during the deposition of the sulfate deposits, the small scattered ephemeral lakes predicted for dry climates could have persisted well into the inferred dry period. Persistent liquid on the surface has implications for life on Mars. These models indicate that if life did exist on Mars surface, it would most likely be found in small localized patches where water could persistently reach the surface during periods of dry climate conditions. This also confirms that impact basins are prime targets for rovers and orbital observation when exploring potential locations for life on Mars, in particular, craters located at the dichotomy boundary.

Although a lake depositional environment and the transition to drier conditions can be inferred from the observed sediment layers in Gale, the climate necessary to form a lake and the influence of lake infill cannot. Since the subsurface, surface, and atmosphere all influence the formation of lakes in Gale, hydrological models are necessary to investigate the past conditions at Gale. A consistent picture of semi-arid conditions when lakes were present in Gale is coming
to light, and further observations of aqueous environments and hydrological models will further constrain the past climates of Mars.
CHAPTER 5

CONCLUSIONS

This work has showed that the consideration of the subsurface, surface, and atmospheric components in the hydrologic cycle is key when investigating fluvial dissection, lake hydrology, and aqueous environments on other planets, just as it is for Earth. While previous work primarily focused on one or two of these components when studying aqueous systems (e.g., Howard, 2007; Andrews-Hanna et al., 2007; Hayes et al., 2008; Hoke et al., 2011), in this thesis I have developed a hydrological model that includes each component of the hydrologic cycle and can be used to study lake hydrology and aqueous environments on extra-terrestrial bodies.

In Chapter 2, the hydrocarbon based hydrology of Titan was explored for both individual lakes and the north polar lake region. At the local scale, results tend to favor high to intermediate permeability aquifers for Titan lakes based on lake stability. The stability of lakes connected to a subsurface aquifer is in agreement with the lack of shoreline change observed in north polar lakes. Furthermore, this stability is in agreement with models that favor large-scale changes in the climate in order to explain, the abundance of dry lakebed features on the surface of Titan. At the regional scale, results favor an aquifer at an intermediate permeability or a high permeability aquifer unit surrounding the pole enclosed by a lower permeability aquifer at lower latitudes, due to the precipitation induced hydraulic head that forces equatorward subsurface flow. A high to low permeability model may be supported by the observation of a radar bright unit surrounding the north polar region (Turtle et al. 2013), and the abundance of karsts in this radar bright unit (Mitchell et al., 2008). For all models, a reduced evaporation rate over the largest sea, Kraken Mare, is required as suggested by others (e.g., Lorenz, 2014), due to the dramatic decrease in evaporation potential at lower latitudes and the areal extent of Kraken Mare. Moreover, these
results showed that subsurface flow is a critical component to the formation and stability of lakes on Titan. Regional subsurface flow at the north polar region removes methane from the wet high latitudes, allowing ponding in topographic depressions at the dry lower latitudes, while local subsurface flow stabilizes lakes during the long dry seasons (~15 Earth-years) on Titan predicting little shoreline change over a Titan year in agreement with observations (Hayes et al., 2012).

The hydrological model developed in Chapter 2 is adapted to Mars in Chapters 3 and 4. In these Chapters, I focus on the hydrology of a paleo-lake in Gale Crater during two distinct periods of lake formation. First I investigate the climate at Gale Crater using lake stands inferred from fan deposits in the early to late Hesperian after the central sediment mound was deposited. While several climate conditions, ranging from arid to semi-arid, can match the observed lake stands in Gale Crater, mineralogical evidence (Ehlmann and Buz, 2016) and evidence for crater lakes outside of Gale (Cabrol and Grin, 1999; Fassett and Head, 2008) are more consistent with a semi-arid climate (aridity index between 3 and 6) with an annual precipitation between 100 and 600 mm/yr for a megaregolith aquifer. This climate range is comparable to steppe climates on Earth for the lowest aridity indices and high altitude, temperate desert climates for the highest aridity indices. This work highlights the complicated interplay between the subsurface, surface, and atmosphere at Gale crater, and emphasizes the need for multiple aqueous indicators when investigating the climate history. While debate over a warm-wet or a cold-icy scenarios for past Mars continues, this work showed that the contribution of subsurface flow to a Gale crater lake in a vertical integrated hydrologic system is non-trivial (between 40% and 50% of the total lake influx). The cold-icy scenario for past Mars, invokes runoff of snow- and ice- melt in the southern highlands to explain the observed channels and lake deposits (e.g., Head et al., 2014).
In this scenario, all of the water that reaches a Gale crater lake must come from overland flow, necessitating large-scale melting of a highlands ice sheet and large volumes of periodic surface discharge. Under the warm and wet scenario, this model illustrated that Gale crater can tap an extensive recharge zone due to its unique location on the dichotomy boundary and thus is less dependent on overland flow for the formation of a Gale crater lake. Furthermore, in climates where precipitation is limited, similar to mid-latitude deserts on Earth, subsurface flow plays a more significant role in the hydrology of a Gale crater lake, suggesting that even under water limited conditions sustained subsurface flow is still predicted at Gale crater.

In Chapter 4, a similar inference of the past climate is determined using the hydrological model, in this case the early history of Gale crater in the late Noachian and early Hesperian. The central sediment mound, Aeolis Mons, was not present during the earliest lake formation in Gale, thus a topographic model is developed in order to reconstruct the topography of a pre-Aeolis Mons Gale crater. Mudstone layers at the base of Aeolis Mons suggest a past lake depositional setting providing constraints on the past climate, requiring conditions wetter than an arid climate ($\phi \leq 6$). The influence of sediment deposition in the central mound was also explored and revealed that the unique location of Gale allows water to preferentially pond in the crater floor even under conditions of thick sediment infill and dry, hyper-arid climate. This suggests that a dramatic climate change, possibly necessitating the cessation of precipitation, is required in order to account for the observed change in depositional environment in the upper layers of the sediment mound. The presence of lakes during sulfate formation, though, cannot be ruled out and may have important implications for potential locations for salts and life on Mars. The results from this Chapter imply that, while sparse, lakes can potentially form in shallow scattered local depressions under hyper-arid conditions and may persist during dry conditions. These localized
depressions may be prime locations to investigate the possibility of past life on Mars. Chloride salt deposits in the vicinity of Gale crater are concentrated in local depressions (Osterloo et al., 2010; Ehlmann and Buz, 2016) similar to the distribution of small shallow lakes predicted under arid to hyper-arid conditions in this model. Future work will investigate the preferential locations of lakes in dry climates and compare to the observed locations of salt deposits.

Although the Cassini mission is nearing its lifetime, plenty of questions remain regarding Titan’s hydrocarbon-based hydrological cycle, including the hydrology and formation of karst landscapes, and the hydrology of lakes in the south polar region. On Mars, Curiosity has just begun its traverse up Aeolis Mons, and the prospect of Mars 2020 landing at either the Jezero Crater delta environment or Mawrth Vallis outflow channel, will provide further constraints on the past aqueous environments on Mars and new locations to investigate with the hydrological model developed in this thesis.


