Earth as an Exoplanet

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The search for habitable and inhabited environments beyond our planet commonly focuses on analogs to Earth, especially in the case of exoplanets. Observations from ground-based facilities, satellites, and spacecraft have yielded a rich collection of data that can be used to effectively view a distant Earth within the context of exoplanet characterization. Application of planetary and exoplanetary remote sensing techniques to these datasets then enables the development of approaches to detecting signatures of habitability and life on other worlds. In addition, an array of models have also been used to simulate exoplanet-like datasets for the distant Earth, thereby providing insights that are often complementary to those from existing observations. Of course, Earth’s atmosphere and surface environment has evolved substantially in the 4.5 billion years since our planet formed. A combination of in situ geological and bio-geochemical modeling studies of our planet have provided glimpses of environments that, while technically belonging to our Earth, are seemingly alien worlds. Understanding the myriad ways Earth has been habitable and inhabited, coupled with remote sensing approaches honed on the distant Earth, provides a key guide to recognizing potentially life-bearing environments in distant planetary systems.

1. INTRODUCTION

The quest for both habitable and inhabited worlds beyond Earth is key to our understanding of the potential distribution of life in the Universe. This ongoing search seeks to answer profound questions: Are we alone? How unique is our Earth? Should the hunt for life beyond Earth uncover a multitude of habitable worlds and few (if any) inhabited ones, humanity would begin to understand just how lonely and fragile our situation is. On the other hand, if our hunt yields a true diversity of inhabited worlds, then we would learn something fundamental about the tenacity of life in the cosmos.

But how will we recognize a distant habitable world, and how would we know if this environment hosts some form of life? A key opportunity for understanding the remote characterization of habitability and life comes from studying our own planet — Earth will always be our best example of a habitable and inhabited world. Thus, by studying our planet within the context of exoplanet exploration and characterization, we develop ideas, approaches, and tools suitable for remotely detecting the signs of habitability and life. While habitable exoplanets planets are unlikely to look exactly like Earth, these worlds will probably share some important characteristics with our own, including the presence of oceans, clouds, surface inhomogeneities, and, potentially, life. Studying globally-averaged observations of Earth within the context of remote sensing therefore provides insights into the ideal measurements to identify planetary habitability from data-limited exoplanet observations.

Of course, Earth is not a static environment. Life emerged on our planet into an environment completely unlike the Earth we understand today. The subsequent evolution of our planet — an intimate coupling between life and geophysical processes — produced worlds seemingly alien to modern Earth. Ranging from ice-covered “Snowball Earth” scenarios to the hazy, oxygen-free atmosphere of the Archean, each evolutionary stage of our planet offers a unique opportunity to understand habitable, life-bearing worlds distinct from the present Earth.

The chapter below summarizes studies of Earth within the context of exoplanet characterization. Following a brief synopsis of the current state of exoplanet science, we discuss the initial experiments that studied our Earth from the perspective of Solar System planetary exploration. Subsequent sections then present both observational and modeling approaches to studying Earth from a distance — the Pale Blue Dot. We then broaden the perspective, and review our understanding of the evolution of Earth, and its appearance, over the last four billion years. Lastly, we present an overview of what has been learned by studying Earth as an exoplanet, summarizing approaches to remote characterization of potentially habitable or inhabited worlds. For further reading, we note that an entire book on studies of the distant Earth has been published by [Vázquez et al. (2010)].

1.1. Current State of Exoplanet Science

Following the first detection of an exoplanet around a Sun-like star (Mayor & Queloz [1995]) and of an exoplanet atmosphere (Charbonneau et al. [2002]) over a decade ago, the field of exoplanetary science has been marked by two clear trends — the steady discovery of increasingly smaller worlds on longer-period orbits, and the ever-increasing quality of observational data suitable for char-
characterizing worlds around other stars. Due to advances in exoplanet detection using a variety of techniques, we now know that, on average, every star in the Milky Way galaxy hosts at least one exoplanet (Casser et al. 2012). Furthermore, due in large part to the success of the Kepler mission (Borucki et al. 2010), we understand that occurrence rates for potentially Earth-like worlds orbiting within the Habitable Zone of Sun-like stars are relatively large, with estimates spanning roughly 10–50% (Dressing & Charbonneau 2013, Petigura et al. 2013, Batalha 2014, Burke et al. 2015, Kopparapu et al. 2018). Excitingly, and especially for low-mass stellar hosts, surveys have revealed a number of nearby potentially Earth-like exoplanets, such as Proxima Centauri b (Anglada-Escude et al. 2016) or the worlds in the TRAPPIST-1 system (Gillon et al. 2016).

The subsequent characterization of exoplanet atmospheres has largely been accomplished using transit and/or secondary eclipse spectroscopy (for a review, see Kreidberg 2018). The former relies on the wavelength-dependent transmittance of an exoplanet atmosphere (Seager & Sasselov 2000, Brown 2001, Hubbard et al. 2001), which causes a transiting world to block more (for lower transmittance) or less (for higher transmittance) light when crossing the disk of its host. By comparison, secondary eclipse spectroscopy measures the planet-to-star flux ratio by observing the combined star and exoplanet spectrum prior to the planet disappearing behind its host star (i.e., secondary eclipse). Using these techniques, astronomers have probed the atmospheres of a striking variety of exoplanets, spanning so-called hot Jupiters (Grillmair et al. 2008, Swain et al. 2008, Pont et al. 2008, Swain et al. 2009, Sing et al. 2009, Madhusudhan & Seager 2009), as well as mini-Neptunes and super-Earths (Stevenson et al. 2010, Bean et al. 2010, Line et al. 2013, Kreidberg et al. 2014, Knutson et al. 2014a,b, Ehrenreich et al. 2014, Fraine et al. 2014, Sing et al. 2016, Stevenson 2016, de Wit et al. 2018). Although, and as with any young field, some findings related to exoplanet atmospheres remain controversial or have undergone substantial revision (Line et al. 2014, Hansen et al. 2014, Diamond-Lowe et al. 2014).

Unfortunately, transit or secondary eclipse spectroscopy is not well-suited to studying the atmospheres of potentially Earth-like planets orbiting within the Habitable Zone of their Sun-like hosts, due to the long orbital periods and low signal sizes for such worlds. Here, direct (or high-contrast) imaging will likely be the leading observational approach, and, as a result, the material below focuses on directly observing Earth in reflected light and/or thermal emission. For exoplanets, direct imaging involves blocking the light of a bright central host star in order to resolve and observe faint companions to that star (Traub & Oppenheimer 2010). Both internal and external occulting technologies are under active study (Guyon et al. 2006, Cash et al. 2007, Shaklan et al. 2010, Mawet et al. 2012), and ground-based telescopes equipped with coronagraphs already enable the characterization of hot gas giant exoplanets orbiting young, nearby stars (Marois et al. 2008, Skemer et al. 2012, Macintosh et al. 2015).

1.2. The Future of Rocky Exoplanets

A number of planned or under-study missions will improve and expand our ability to characterize exoplanet atmospheres and surfaces. Foremost among these is NASA’s James Webb Space Telescope (JWST) (Gardner et al. 2006), which is expected to provide high-quality transit and secondary eclipse spectra of many tens of targets over the duration its designed five year mission (Beichman et al. 2014). Some of these observations will probe lower-mass, potentially rocky exoplanets (Deming et al. 2009, Batalha et al. 2015). Critically, JWST may even be capable of characterizing temperate Earth-sized planets orbiting low-mass stars (Kaltenegger & Traub 2009, Cowan et al. 2015, Barstow et al. 2016), though the ability to conduct such studies depends largely on the behavior and size of systematic noise sources (Greene et al. 2016).

Following JWST, NASA will launch the Wide-Field InfraRed Survey Telescope (WFIRST: Spengel et al. 2013). It is anticipated that WFIRST will be equipped with a Coronagraphic Instrument (CGI) capable of visible-light imaging and, potentially, spectroscopy of exoplanets (Noecker et al. 2016). Key outcomes of this mission will include a demonstration of high-precision coronagraphy in space, as well as the study of a small handful of cool, gas giant exoplanets (Marley et al. 2014, Hu 2014, Burrows 2014, Lupu et al. 2016, Traub et al. 2016, Nayak et al. 2017). However, the planned capabilities of WFIRST/CGI will make observations of Earth-like planets extremely unlikely (Robinson et al. 2016).

Exoplanet direct imaging missions that could build on the technological successes of WFIRST are already under investigation. Included here are the WFIRST starshade rendezvous concept (Seager et al. 2015), the Habitable Exoplanet (HabEx) imaging mission (Mennesson et al. 2016), and the Large Ultraviolet-Visible-InfraRed (LUVIOR) explorer (Peterson et al. 2017). While the scope and capabilities of these under-study missions are varied (Stark et al. 2016), a central goal unites these concepts — to detect and characterize Pale Blue Dots around our nearest stellar neighbors.

2. EARTH AS A PLANET

The Pioneer 10/11 (launched 1972; Baker et al. 1975), Voyager 1/2 (launched 1977; Kohlhase & Penzo 1977, Hanel et al. 1977) enabled the initial development of key approaches to analyzing spacecraft flyby data for many Solar System worlds. Instruments and techniques for either acquiring or interpreting spatially-resolved observations of planets and moons using pho-
ometry or spectroscopy at wavelengths spanning the ultraviolet through the infrared were among the important developments. From these observations planetary scientists were able to infer details about atmospheric chemistry and composition, cloud and aerosol formation and distribution, atmospheric thermal structure and circulation, surface chemical and thermal properties (for worlds with thin atmospheres), as well as planetary energy balance.

Flybys of Earth by the Galileo spacecraft (launched 1989; Johnson et al. 1992) in December of 1990 and 1992 afforded planetary scientists the first opportunity to analyze our planet using the same tools and techniques that had been (and would be) applied throughout the Solar System. During the flybys, data were acquired using the Near-Infrared Mapping Spectrometer (NIMS; Carlson et al. 1992), Solid-State Imaging system (SSI; Belton et al. 1992), Ultraviolet Spectrometer (UVS; Hord et al. 1992), and the Plasma Wave Subsystem (PWS; Gurnett et al. 1992). Critically, spatially-resolved imagery from the SSI was provided across eight filters (Belton et al. 1992) and spatially-resolved spectra were acquired by NIMS over 0.7–5.2 µm (at a resolution of 0.025 µm longward of 1 µm, and 0.0125 µm shortward of 1 µm).

Figure 1 shows a sampling of SSI images of Earth from the first Earth flyby and a time sequence of SSI images of Earth and the Moon that were acquired after the second Earth flyby. Many images from the second Earth flyby suffered from saturation defects. Reconstructed NIMS images from both Earth flybys are shown in Figure 2 where certain instrument and mapping defects can be seen. Here, the 4.0 µm images highlight a wavelength range with relatively little atmospheric opacity and where thermal emission dominates. By contrast, the 2.75 µm image (located within a H$_2$O absorption band) contains both reflected and thermal contributions, which, for example, results in reflective clouds only being seen in the sunlit portions of the images.

In a landmark study, Sagan et al. (1993) used the Galileo flyby data to ascertain key details about the surface and atmospheric state of our planet. Spectra from the NIMS instrument contain information about surface and atmospheric chemistry, and can also indicate surface thermal conditions at longer wavelengths. Thus, Sagan et al. (1993) argued that reflective polar caps seen in SSI images were water ice, and that the surface of the planet spanned the freezing point of water (covering at least 240–290 K). Additionally, the darkest regions in the SSI images showed signs of specular reflection, indicating that these regions were liquid oceans. Finally, clearsky soundings of H$_2$O absorption features indicated a surface with large relative humidity (i.e., near the condensation point). Taken altogether, these lines of evidence clearly indicate that the world under investigation is habitable, or capable of maintaining liquid water on its surface.

Drossart et al. (1993) retrieved abundances of key atmospheric constituents (CO$_2$, H$_2$O, CO, O$_3$, CH$_4$, and N$_2$O) via simple parameterized fits to resolved NIMS observations (see Figure 3). Here, model spectra were generated by adopting scaled Earth-like profiles for trace atmospheric species. Thermal structure profiles were derived using the 4.3 µm CO$_2$ band, which rely on this gas having a well-mixed vertical profile (i.e., a near-constant mixing ratio with altitude). For a well-mixed gas, variation in an infrared absorption band can be attributed to thermal structure rather than variation in abundance with altitude.

Observations from the NIMS instrument also indicated large column densities of O$_2$ in Earth’s atmosphere (Sagan et al. 1993). It was estimated that diffusion-limited escape of hydrogen (produced from H$_2$O photolysis) would require many billions of years to build up atmospheric oxygen to the observed levels, implying an alternative source. Using the abundances derived in Drossart et al. (1993), it was shown that atmospheric CH$_4$ was in a state of extreme disequilibrium with no known geological source that could supply CH$_4$ at the rate required to maintain the observed concentrations. Certain surface regions of the planet imaged in multiple SSI filters demonstrated a sharp increase in reflectivity at wavelengths beyond 700 nm, known from ground-truth investigations to be the vegetation “red edge.” Thus, multiple lines of evidence point towards the potential for biological activity to be shaping the surface and atmospheric properties of Earth.

Of course, the strongest evidence for Earth being inhabited came from the PWS dataset (Sagan et al. 1993). Here, radio emissions from our planet were monitored as a function of time and frequency throughout the Galileo encounter. Narrow-band emissions between 4–5 MHz (i.e., in the high frequency portion of the radio spectrum where a variety of radio communications occur), isolated in both frequency and time, were interpreted as radio transmissions from an intelligent species on our planet. In other words, an observational approach championed in the search for extraterrestrial intelligence (SETI) — listening at radio frequencies — yielded the best evidence for the inhabitance of Earth from the Galileo flyby dataset.

In the words of Sagan et al. (1993), the Galileo Earth datasets offered a “unique control experiment on the ability of flyby spacecraft to detect life at various stages of evolutionary development.” Combining lines of evidence that spanned the ultraviolet, visible, infrared, and radio spectral regimes, the Galileo observations indicated a habitable planet with a diversity of surface environments and whose atmosphere (and, thus, spectrum) is strongly influenced by life. In the context of exoplanets, however, the key question becomes: Which habitability and life signatures are lost when Earth is studied not as a resolved source but as a distant, unresolved target?

3. OBSERVING THE DISTANT EARTH

The resolution of a telescope is limited by the physics of light diffraction to an angular size of roughly $\lambda/D$, where $\lambda$
Fig. 1.— Top: Time sequence of Earth and the Moon acquired with the SSI “infrared” filter (961–1011 nm) after the second Galileo flyby (December, 1992). Bottom: Images of Earth acquired with the SSI from the first Galileo flyby (December, 1990). Filters from left to right are violet (382–427 nm), green (527–592 nm), red (641–701 nm), and “infrared”. The sub-observer point in each image is approximately identical, and Australia is the landmass near the center of the images. For all images south is oriented “up,” as is in the true flyby geometry.

Fig. 2.— Reconstructed images of Earth using NIMS observations from the first (top) and second (bottom) Galileo Earth flybys. Images in the left column are at 2.75 µm while images in the right column are at 4.0 µm. The bright (warm) source seen in both 4.0 µm images is likely Australia. The spacecraft was nearer to Earth in the second flyby dataset, thereby providing better resolution across the disk.
is wavelength and $D$ is telescope diameter. For a 10-meter class telescope observing at visible wavelengths (i.e., near 500 nm), the angular resolution is at best $5 \times 10^{-8}$ radians (or about 10 milli-arcseconds). Even for our nearest stellar neighbors (e.g., α Centauri at 1.3 parsecs) this corresponds to a spatial resolution of $2 \times 10^{6}$ km, or about three times the radius of our Sun. Thus, even far-future telescopes will not be able to resolve the surfaces of exoplanets (although time-domain observations can be used to obtain some spatial resolution; e.g., Cowan et al. 2009; Majeau et al. 2012) — the better model for future observations of Earth-like exoplanets are not the Galileo flyby observations, but instead the famous “Pale Blue Dot” photograph of Earth taken by the Voyager 1 spacecraft (see Figure 4).

Unresolved observations of planets are sometimes referred to as being “disk-integrated.” Here, the entire three-dimensional complexity of a planet is effectively collapsed into a single pixel. For worlds like Earth, this means that cloud-free regions of the planet are blended with cloudy regions, warm equatorial zones are observationally mixed with cold polar caps, and continents become unresolved from oceans. Additionally, viewing geometry plays an important role, as the portions of the planet near the limb will contribute less overall flux to an observation (since these areas are of smaller solid angular size), and, in reflected light, regions near the day/night terminator will also contribute relatively little flux owing to their lower insololation.

If we wish to extend the analysis techniques applied to the Galileo Earth flyby observations to an unresolved Pale Blue Dot, we must look towards either observational datasets for an unresolved Earth, or towards datasets or products that can mimic an unresolved Earth. There are, in general, three approaches to obtaining (or constructing) such disk-integrated observations of our planet. First, one can observe light reflected from Earth from the portion of the Moon that is not illuminated by the Sun (i.e., so-called “Earthshine” observations; Danjon 1928; Dubois 1947; Woolf et al. 2002; Pallé et al. 2003; Turnbull et al. 2006). Second, one can use spatially resolved observations from satellites in Earth orbit to piece together a disk-integrated view of Earth (Hearty et al. 2009). Finally, spacecraft observations of the distant Earth — like those acquired by Galileo — can be integrated over the planetary disk to yield exoplanet-like datasets. We discuss each of these approaches below, highlighting the advantages and disadvantages of each technique.

### 3.1. Earthshine

Using the dark portion of the Moon — which is illuminated by Earth but not the Sun — has a long history of revealing key details about our planet. In the Dialogue Concerning the Two Chief World Systems, Galileo used Earthshine to deduce that “seas would appear darker, and [. . .] land brighter” when observed from a distance (Galilei 1632). In the first multi-year Earthshine monitoring experiment, described in Danjon (1928) and continued by Dubois (1947), the broadband visual reflectivity of Earth was shown to vary by several tens of percent at a given phase angle (i.e., the planet-star-observer angle), and cloud variability was identified as the likely driver of these variations.

Modern Earthshine observations (Goode et al. 2001; Woolf et al. 2002; Pallé et al. 2004b) have reached an impressive level of precision. Achieving this precision requires corrections for airmass effects, the lunar phase function, and variations in reflectivity across the lunar surface. Nevertheless, it is now common for Earthshine measurements to achieve 1% precision on a given night (Qiu et al. 2003). Such high-quality photometric observations have revealed variability in the visible reflectivity of Earth at
daily, monthly, seasonal, and decadal timescales (Goode et al. 2001; Pallé et al. 2003, 2004a, 2009a; Palle et al. 2016). Figure 5 shows a collection of phase-dependent visual (400–700 nm) apparent albedo measurements from Earthshine measurements. Apparent albedo ($A_{\text{app}}$) is defined by normalizing an observed planetary flux to that from a perfectly reflecting Lambert sphere observed at the same phase angle, or

$$A_{\text{app}} = \frac{3}{2} \frac{F_p}{F_s} \frac{\pi}{\sin \alpha + (\pi - \alpha) \cos \alpha},$$

(1)

where $\alpha$ is the star-planet-observer (i.e., phase) angle, $F_p$ is the planetary flux scaled to the top of the atmosphere, $F_s$ is the solar/stellar flux at normal incidence on the top of the planetary atmosphere, and the flux quantities can either be wavelength-dependent (resulting in a wavelength-dependent apparent albedo) or integrated. Note that the factor of $3/2$ comes from the conversion between geometric and spherical albedo.

Spectroscopic studies of Earthshine (Woolf et al. 2002) offer additional insights into Earth as an exoplanet (Figure 6), beyond those obtained through photometric Earthshine investigations. Using spectroscopic Earthshine data collected over several weeks or months, Arnold et al. (2002) and Seager et al. (2005) showed that the aforementioned vegetation red edge signature is variable in the reflectance spectrum of Earth, and can lead to sharp reflectivity increases at the 10% level in the 600–800 nm range. A red edge-focused Earthshine study by Montañés-Rodríguez et al. (2006) found no strong signature in spectroscopic data from a single night, likely pointing to the importance of cloud cover in masking surface reflectance features. In Earthshine observations that spanned 0.7–2.4 μm, Turnbull et al. (2006) noted a plethora of absorption features that were indicative of life, habitability, and geological activity. Also, after accounting for how the lunar surface de-polarizes radiation, polarization-sensitive spectroscopic Earthshine observations have explored the degree to which a spectrum of Earth can be polarized (0–20%, depending on wavelength) as well as the impact of cloud cover on this signature (Sterzik et al. 2012; Miles-Páez et al. 2014). Finally, by investigating the Earthshine spectrum at extremely high spectral resolution González-Merino et al. (2013) uncovered narrow spectral features due to atomic sodium in Earth’s atmosphere that are either of terrestrial or meteoritic origin.

Of course, Earth-like planets around other stars may not be solely investigated using reflected-light techniques, especially in the case of potentially habitable worlds orbiting M dwarf hosts where transit or secondary eclipse observations would be the preferred approach. Impressively, observational techniques developed for Earthshine data collection have been re-purposed to enable observations of the transmission spectrum of Earth’s atmosphere. By observing the Moon during a lunar eclipse, Pallé et al. (2009b) were able to measure light that had been transmitted through our atmosphere and reflected by the lunar surface. These observations revealed signatures of key atmospheric and biosignature gases, and even included narrow features due to ionized calcium as well as broad pressure-induced features from O$_2$ and N$_2$ (the latter of which is typically difficult to detect due to its general lack of ro-vibrational features). A follow-up analysis of these data by García Muñoz et al. (2012) showed that refractive effects in transit spectra of Earth twins would limit the atmospheric depths probed (during mid-transit) to be above about 10 km, thus providing limited information from the surface and tropospheric environments. Also, additional high-resolution transmission spectra acquired using Earthshine-related techniques have revealed variability in the depths of H$_2$O absorption features (Yan et al. 2015), likely tied to the condensable nature of this gas in our atmosphere.

Finally, while Earthshine techniques have been proven to be both powerful and versatile, this approach does have its shortcomings. First, due to the ground-based nature of
Fig. 5.— Measurements of the phase-dependent visual (400–700 nm) apparent albedo of Earth from Earthshine data spanning several years (from Pallé et al. 2003), and from several spacecraft missions. Apparent albedo values larger than unity indicate stronger directional scattering than can be produced by a Lambert sphere. The DSCOVR datapoint is derived from the available four narrowband channels that span the visible range since an integrated 400–700 nm observation cannot be produced from DSCOVR data.

Fig. 6.— Scaled reflectance spectrum of Earth at visible and near-infrared wavelengths measured from Earthshine. Key absorption and reflection features are indicated. Data courtesy M. Turnbull from Turnbull et al. (2006).
observations, full diurnal cycles in the reflectivity of Earth cannot be observed except during polar night. Second, it is often difficult to calibrate Earthshine observations in a fashion that reveals the absolute brightness of our planet. Thus, some Earthshine datasets are only reported as a scaled reflectance value, and these products are of lower utility when it comes to exoplanet detectability and characterization studies. Finally, Earthshine cannot be used to observe thermal emission from Earth since the Moon is also self-luminous at infrared wavelengths.

3.2. Orbit

A large suite of satellites are continuously monitoring the Earth system from space. While most of these Earth-observing satellites only resolve a small patch of our planet in any individual observation, the collective dataset from these satellites benefit from extensive temporal, spatial, and spectral coverage. Thus “stitching” together spatially-resolved radiance measurements from one (or several) observing platform(s) can enable a view of the entire disk of Earth. This approach was pioneered by Hearty et al. (2009), who used spatially-resolved thermal radiance observations from the Atmospheric Infrared Sounder (AIRS) instrument (aboard NASA’s *Aqua* satellite; Aumann et al. 2003) to create disk-integrated infrared spectra of Earth (Figure 7).

The practice of stitching together resolved radiance measurements from an Earth-observing satellite is, unfortunately, not straightforward. Temporal gaps sometimes exist in these datasets where a given latitude/longitude patch of Earth has not been observed in a given 24 hr period. Thus, if the goal is to produce a snapshot of Earth at a given time, an interpolation of existing radiance observations across time must be performed. As the Earth climate system (as well as top-of-atmosphere radiances) is non-linear, this interpolation introduces some uncertainties.

The greater challenge to deriving whole-Earth views from resolved satellite observations, though, stems from viewing geometry constraints. Most Earth-observing satellites are designed to acquire observations in the nadir (i.e., direct downward) direction. For satellites observing in reflected light, the range of solar incidence angles can also be limited, especially over any given several-day period. Thus, when stitching together a whole-disk observation, data for certain viewing geometries (e.g., patches located near the limb, where the observing geometry is quite distinct from nadir-looking) may not exist. In this case, assumptions must be adopted for how radiance will vary with the emission angle and/or the solar incidence angle. For example, Hearty et al. (2009) adopted a limb darkening law to transform radiances acquired at nadir to radiances appropriate for other emission angles.

An alternative approach to using directly-observed radiances from satellites is, instead, to adopt a collection of satellite-derived “scene” models. These scene models describe the viewing geometry-dependent brightness of different surface categories for Earth (e.g., ocean or desert) under different cloud coverage scenarios. In other words, these models specify the bi-directional reflectance distribution functions for a large variety of surface type and cloud coverage combinations. Scene models can be derived from satellite observations (Suttles et al. 1988) or can be designed to fit satellite observations (Manalo-Smith et al. 1998). Combining data that describe the time-dependent distribution of clouds, snow, and ice on Earth with a set of scene models then enables the recreation of whole-disk views of our planet (Pallé et al. 2003; Oakley & Cash 2009). Integrating these three-dimensional models over the planetary disk then yields the brightness (or reflectivity) of the Pale Blue Dot. One key shortcoming of the scene model approach, however, is that such models are rarely spectrally resolved, and instead specify a broadband reflectivity or brightness. Thus, such models cannot produce spectra of the disk-integrated Earth, and, instead, focus on computing broadband lightcurves for the Pale Blue Dot.

Solar occultation observations acquired from Earth orbit provide a direct measurement of the transmittance along a slant path through the atmosphere. Initially such datasets (Abrams et al. 1999; Bovensmann et al. 1999; Bernath et al. 2005) provided an excellent model validation for tools designed to simulate transit spectra of Earth-like exoplanets (Kaltenegger & Traub 2009; Misra et al. 2014b). However, as was recognized by Robinson et al. (2014) and Dalba et al. (2015), occultation observations orbit can be directly translated into transit spectra. Using data from the Canadian Atmospheric Chemistry Experiment - Fourier Transform Spectrometer (ACE-FTS; Bernath et al. 2005), Schreier et al. (2018) created transit spectra of Earth spanning 2.2–13.3 μm and demonstrated that signatures of chlorofluorocarbons appeared in the occultation-derived transit observations, in addition to more-standard features of H₂O, CO₂, CH₄, N₂O, N₂, NO₂, and O₂.

3.3. Spacecraft

The ideal approach for mimicking direct observations of Earth-like exoplanets is, of course, to acquire photometry and/or spectroscopy for a truly distant Earth. Such observations must be taken from distances beyond low-Earth or geostationary orbit, as the entire disk of the planet is not entirely visible from these vantages. Thus, views of Earth from spacecraft at lunar distances or from Earth-Sun Lagrange points, or observations from interplanetary spacecraft, are all excellent sources. Until the recent launch of the Deep Space Climate Observatory (DSCOVR; Biesecker et al. 2015) mission to the Earth-Sun L1 point, no dedicated mission existed for observing Earth from a great distance. Thus, the majority of the spacecraft observations relevant to Earth as an exoplanet came from missions sent to other Solar System worlds.

While spacecraft observations of the distant Earth are ideal for exoplanet-themed investigations, this approach is
Fig. 7.— Disk-integrated thermal infrared spectra of Earth from the AIRS instrument (Hearty et al. 2009) and from the Mars Global Surveyor Thermal Emission Spectrometer (MGS/TES; Christensen & Pearl 1997). Key features are labeled and blackbody spectra at different emitting temperatures are shown. Inset is a broadband thermal infrared (6–10 µm) image of Earth from the LCROSS mission (Robinson et al. 2014).

not without its shortcomings. First, it is difficult to find time during the main phase of a mission to dedicate towards observations of non-primary targets such as Earth. This means that the temporal coverage of spacecraft datasets for the distant Earth is poor, with many of these datasets acquired during the cruise phase of a mission. Second, and most unfortunately, spacecraft datasets for the distant Earth often remain unpublished. In these circumstances, the data may have been acquired only for press or outreach purposes, or it might be that analysis and publication of these data are seen as a distraction from the main goals of a mission. Unpublished datasets are known to exist for a number of other missions including: Cassini, Clementine, Lunar Reconnaissance Orbiter, Mars Express, Mars Reconnaissance Orbiter, MESSENGER, OSIRIS-REx, SELENE/Kaguya, and Venus Express.

A detailing of published spacecraft-acquired datasets that are relevant to Earth as an exoplanet is shown in Table 1 emphasizing photometric and spectroscopic observations that span the ultraviolet, visible, and infrared wavelengths. Beyond the previously-discussed Galileo Earth flyby observations, key datasets also come from a snapshot thermal infrared spectrum acquired by the Mars Global Surveyor Thermal Emission Spectrometer (MGS/TES), visible photometry and near-infrared spectroscopy spanning 24 hr on five separate dates from the EPOXI mission (which repurposed the Deep Impact flyby spacecraft), visible spectroscopy and infrared photometry and spectroscopy taken over brief intervals on three separate dates by the Lunar Crater Observation and Sensing Satellite (LCROSS), and the aforementioned DSCOVR data (which include images taken in 10 narrowband channels spanning ultraviolet and visible wavelengths, and bolometric measurements in several channels spanning 0.2–100 µm). The EPOXI dataset has been used to analyze key spectral features for Earth in the near-infrared range and to quantify the vegetation red edge signature in disk-integrated observations (Louden et al. 2011; Robinson et al. 2011), and to investigate mapping techniques for unresolved objects (Cowan et al. 2009; Fujii et al. 2011; Cowan et al. 2011). In Robinson et al. (2014), the LCROSS Earth observations were used to quantify the impact of ocean glint and ozone absorption on phase-dependent disk-integrated visible spectroscopic data for the Pale Blue Dot. A digest of ultraviolet, visible, and near-infrared observations is shown in Figure 8.

4. MODELING THE PALE BLUE DOT

Techniques for simulating observations of the distant Earth provide a complementary approach to spacecraft, orbital, and Earthshine observations. Especially once validated against observational datasets, models of the disk-integrated Earth enable the exploration of the Pale Blue Dot across a wide range of wavelengths and spectral resolutions, and can also fill in the various gaps that exist between different observational approaches. Currently, a hierarchy of Earth models exists, spanning simple reflectance tools to complex three-dimensional models whose outputs cover the ultraviolet through the far-infrared.

4.1. One-Dimensional Approaches

One-dimensional models of the Pale Blue Dot capture the vertical structure of Earth’s atmosphere, but omit any latitudinal or longitudinal structure in the atmosphere and surface. Such simplifications enable these one-dimensional approaches to be computationally efficient, and often allow for higher spectral resolution in model outputs. Nevertheless, key details about the fractional distribution of clouds and various surface types on Earth must be accounted for,
Table 1
PUBLISHED SPACECRAFT DATASETS FOR EARTH AS AN EXOPLANET

| Spacecraft | Date   | Phase Angle(s) | Source(s)                                                                 |
|------------|--------|----------------|---------------------------------------------------------------------------|
| Galileo    | 1990-12-10 | 35°             |                                                                            |
|            | 1992-12-09 | 82°             | Sagan et al. (1993); Drossart et al. (1993)                               |
|            | 1992-12-16 | 89°             |                                                                            |
| MGS/TES    | 1996-11-23 | n/a             | Christensen & Pearl (1997)                                               |
|            | 2008-03-18 | 58°             |                                                                            |
|            | 2008-05-28 | 75°             |                                                                            |
| EPOXI      | 2008-06-04 | 77°             | Livengood et al. (2011); Cowan et al. (2011); Fujii et al. (2011); Robinson et al. (2011) |
|            | 2009-03-27 | 87°             |                                                                            |
|            | 2009-10-04 | 86°             |                                                                            |
|            | 2009-08-01 | 23°             |                                                                            |
| LCROSS     | 2009-08-17 | 129°            | Robinson et al. (2014)                                                    |
|            | 2009-09-18 | 75°             |                                                                            |
| DSCOVR     | ongoing  | 4–12°           | Biesecker et al. (2015); Yang et al. (2018)                               |

*a*Based on UT at start of observations.

Fig. 8.—Summary of published observations of the distant Earth at ultraviolet, visible, and near-infrared wavelengths. The first figure presents spectra of Earth’s apparent albedo from Galileo, EPOXI, and LCROSS. A crescent-phase observation from LCROSS is marked by large apparent albedo, which was driven primarily by forward scattering from a well-defined glint spot. The second figure presents near-infrared emission observations from Galileo and EPOXI, and blackbody spectra are provided. Key absorption features are indicated in both figures.
either through data-informed weighting factors or through tuning parameters.

Traub & Jucks (2002) presented one of the earliest models of the Pale Blue Dot. This one-dimensional tool spanned the ultraviolet through thermal infrared, and included absorption and emission from key atmospheric species. Radiation multiple scattering was neglected, and modeled observations in reflected light were generated by linearly combining spectral components (including Rayleigh, clear sky, high cloud, and others). At visible wavelengths, disk-integrated observations were simulated using a single solar zenith angle (i.e., the Sun was placed at a zenith angle of 60° over a plane-parallel atmosphere). Both Woolf et al. (2002) and Turnbull et al. (2006) used the Traub & Jucks (2002) model to analyze Earthshine spectra. By fitting the reflected-light spectral components in the Traub & Jucks (2002) model to the Earthshine data, these authors determined that the most important aspects of their reflected-light observations were the Rayleigh component, a gray high cloud continuum, and Rayleigh scattering. More recently, the Traub & Jucks (2002) model has been used to study the spectral evolution of Earth through time (Kaltenegger et al. 2007) (Rugheimer & Kaltenegger 2018), including a comparison to the previously mentioned EPOXI dataset (Rugheimer et al. 2013).

A multiple-scattering one-dimensional model, developed by Martín-Torres et al. (2003), was adopted by Monta˜n´es-Rodr´ıguez et al. (2006) to help understand the signature of the vegetation red edge in Earthshine spectra. In this work, the strong match between the Earthshine data and the simulations was attributed to the scattering treatment within the model. Also, the Monta˜n´es-Rodr´ıguez et al. (2006) study developed a sophisticated approach to capturing the latitudinal and longitudinal distribution of clouds and surface types on Earth. Specifically, disk-averaged cloud and surface coverage maps were derived from Earth science data products, including appropriate weighting factors for the solar and lunar geometry.

4.2. Three-Dimensional Models

In general, three-dimensional models of the Pale Blue Dot compute the spatially-resolved radiance over the planetary disk, and then integrate this radiance over solid angle to produce a disk-integrated quantity. More formally, three-dimensional models of Earth aim to compute the integral of the projected area weighted intensity in the direction of an observer, which is written as,

\[ F_\lambda (\hat{o}, \hat{s}) = \frac{R_E^2}{d^2} \int_{2\pi} I_\lambda (\hat{n}, \hat{o}, \hat{s}) (\hat{n} \cdot \hat{o}) \, d\omega, \tag{2} \]

where \( F_\lambda \) is the disk-integrated specific flux density received from a world of radius \( R_E \) at a distance \( d \) from the observer, \( I_\lambda (\hat{n}, \hat{o}, \hat{s}) \) is the location-dependent specific intensity in the direction of the observer, \( d\omega \) is an infinitesimally small unit of solid angle on the globe, \( \hat{n} \) is a surface normal unit vector for the portion of the surface corresponding to \( d\omega \), and \( \hat{o} \) and \( \hat{s} \) are unit vectors in the direction of the observer and the Sun, respectively (see Figure 9). The integral in Equation 2 is over the entire observable hemisphere (2\( \pi \) steradians) and the dot product at the end of the expression ensures that an element of area \( R_E^2 \, d\omega \) near the limb is weighted less than an element of equal size near the sub-observer point. Note that, for reflected light, \( I_\lambda \) will be zero at locations on the night side of the world (i.e., where \( \hat{n} \cdot \hat{s} < 0 \)), but is non-zero at all locations when considering thermal emission.

The most straightforward three-dimensional models use empirical bi-directional reflectance distribution functions (e.g., the previously-mentioned scene models from Manalo-Smith et al. 1998) to specify the reflectivity of a patch on the disk as a function of viewing geometry. These three-dimensional tools can either be spectrally-resolved (Ford et al. 2001) or broadband (McCullough 2006; Pallé et al. 2003, 2008; Williams & Gaidos 2008; Oakley & Cash 2009). Atmospheric effects (e.g., gas absorption and scattering) are typically omitted, although Fuji et al. (2010) produced a three-dimensional reflectance model that blended wavelength-dependent bi-directional reflectance distribution functions from a variety of sources and also included an additive atmospheric Rayleigh scattering term. Time-dependent distributions of clouds and surface types are derived from Earth science datasets, such as the International Satellite Cloud Climatology Project (ISCCP; Schiffer & Rossow 1983).

The most complex three-dimensional tools for simulating observations of the distant Earth solve the full plane-parallel, multiple-scattering radiative transfer equation to determine the emergent radiance over the planetary disk. By including realistic atmospheric radiative effects, these fully multiple-scattering tools can more self-consistently capture gas and cloud absorption and scattering (Tinetti et al. 2006; Fuji et al. 2011; Robinson et al. 2011; Feng et al. 2018) as well as polarization effects (Stam 2008). Like the previously-discussed reflectance models, cloud and surface type coverages are typically derived from Earth science datasets, while cloud optical thicknesses must also be adopted from similar datasets to include in the multiple-scattering calculation. Such sophisticated three-dimensional models can serve as virtual “laboratories” for studying the Pale Blue Dot across a wide range of timescales, wavelengths, and viewing geometries (see Figure 10).

5. PALEO-EARTH AS AN EXOPLANET

Earth has evolved dramatically over time, with the physicochemical properties of Earth’s interior, crust, and ocean-atmosphere system, as well as the mechanisms coupling these components together, undergoing significant changes across a wide range of timescales. These changes
Fig. 9.— Geometry for modeling disk-integrated Earth observations. The surface normal vector, and the vectors in the direction of the observer and Sun are $\hat{n}$, $\hat{o}$, and $\hat{s}$, respectively. The angle $\alpha$ is the phase angle, while $\phi$ and $\theta$ are the coordinates of latitude and longitude, respectively. Earth view generated by the Earth and Moon Viewer, first implemented by J. Walker (http://www.fourmilab.ch/cgi-bin/Earth).

Fig. 10.— Simulations of Earth’s phase- and wavelength-dependent apparent albedo (from Robinson et al. 2010). Models are averaged over a full rotation at each phase, and the angles of the given phases are 0° (full), 45° (gibbous), 90° (quadrature), and 135° (crescent). Large apparent albedos at crescent phase are primarily due to ocean glint and cloud forward scattering. The similarity in apparent albedo scales for the quadrature, gibbous, and full spectra indicate that Earth largely scatters like a Lambert sphere across these phase angles. A slight enhancement in apparent albedo at full phase is due to cloud back scattering.
have in turn impacted both the habitability of Earth surface environments (e.g., Kasting & Catling 2003) and the remote detectability of Earth’s biosphere (Kaltenegger et al. 2007; Meadows 2008; Reinhard et al. 2017a; Rugheimer & Kaltenegger 2018). In particular, the atmospheric abundances of almost all potential biosignature gases (e.g., CH$_4$, O$_2$, O$_3$, N$_2$O, CO$_2$) have changed by many orders of magnitude throughout Earth’s history. The timing and magnitude of these changes have been controlled by often complex interactions between biological, geologic, and stellar factors. At the same time, Earth’s climate system and surface habitability have changed significantly, as influenced by both long-term trends in stellar energy flux, catastrophic climate destabilization during low-latitude “Snowball Earth” glaciations, and major impact events.

Despite these dramatic changes, all life on Earth appears to share a single phylogenetic origin that is perhaps nearly as ancient as Earth itself (Fox et al. 1980). Earth’s history thus allows us to explore the long-term evolutionary factors controlling the production and maintenance of remotely detectable signatures of habitability and life against the backdrop of a continuously inhabited planet. Fully illuminating this history requires integration of geologic and geochemical data with quantitative mechanistic models, with the ultimate goal of giving such models predictive power in the search for living planets beyond Earth.

Here, our focus is on observations and models aimed at constraining surface habitability and atmospheric composition through time on Earth, with a particular eye toward the detectability of habitability markers and atmospheric biosignatures. We do not explicitly evaluate the potential roles of mantle redox (Kump et al. 2001; Li & Lee 2004; Nicklas et al. 2018), crustal differentiation (Kump & Barley 2007; Tang et al. 2016b; Greber et al. 2017), or tectonic mode (Korenaga 2013), but emphasize that these factors must also be fully incorporated into any predictive model for exoplanet habitability and life detection.

5.1. Geological Constraints

With some notable exceptions (see below), Earth’s climate appears to have been central for the vast majority of its history. Oxygen and lithium isotope evidence from ancient zircons suggest the presence of liquid water at Earth’s surface by 4.3 billion years ago (Ga) (Mojzsis et al. 2001; Ushikubo et al. 2008), and a consistent if fragmentary marine sedimentary rock record attests to a large-scale fluid-mediated rock cycle for the last 3.8 billion years (e.g., Rosing et al. 1996). The oxygen isotope composition of marine cherts (Knauth & Epstein 1976; Robert & Chaussidon 2006) and the temperature stability of reconstructed ancestral proteins (Gaucher et al. 2008) have been used to suggest that surface temperatures during much of the Archean Eon (3.8–2.5 Ga) were hot, perhaps as high as 70°C. More recent estimates based on the oxygen and hydrogen isotope composition of cherts (Hren et al. 2009) and the oxygen isotope composition of putatively biogenic phosphate minerals (Blake et al. 2010) suggest much cooler (but still quite warm) temperatures between 25–40°C.

Striated clasts and ice-rafted debris are observed in diamictites of the Mozaan Group, South Africa at ∼2.9 Ga (Young et al. 1998), representing the most well-established glacial deposits of Archean age. Given existing paleolatitude constraints of around 45–50° (Kopp et al. 2005), these deposits suggest a climate similar to or colder than that of the Pleistocene Earth. More recently, diamictites and rhythms containing dropstones have been reported from the ∼3.5 Ga Overwacht Group, Barberton Greenstone Belt, South Africa (de Wit & Furnes 2016). These strata have also been interpreted as volcanic breccias (e.g., Viljoen & Viljoen 1969), and there is a conspicuous absence of striated clasts. However, the reconstructed paleolatitudes of these deposits are between ∼20–40°, so if they are indeed glaciogenic in origin they would imply a relatively cold early Archean climate.

Firm evidence for glaciation does not reappear until after the Archean-Proterozoic boundary, with a series of glacial deposits observed in North America (Young 2001) and South Africa (Rasmussen et al. 2013) between ∼2.4 – 2.3 Ga. Glaciogenic deposits found on the Kaapval craton, South Africa, recently dated to 2.426±0.003 Ga (Gumsley et al. 2017), show evidence for being deposited at low latitudes (Evans et al. 1997), leading to the suggestion that these deposits record a Paleoproterozoic “Snowball Earth” — classically envisaged as a catastrophic destabilization of the climate system during which runaway ice-albedo feedback causes the advance of ice sheets to the tropics and a virtual shutdown of the hydrologic cycle (Budyko 1969; Sellers 1969). Indeed, the temporal correspondence between this apparently intense ice age and the initial accumulation of O$_2$ in Earth’s atmosphere has led to the suggestion that the climate system was transiently destabilized by a sharp drop in atmospheric CH$_4$ attendant to rising atmospheric O$_2$ (Kasting 2005).

Recent evidence suggests that these intense ice ages were followed by a transient period of atmospheric oxygenation, after which the climate system appears to have been relatively stable with little firm evidence for glaciation between ∼1.8 – 0.8 billion years ago (here referred to as the ‘mid-Proterozoic’). A notable exception to this comes in the form of putative glacial deposits from the Vazante Group, east-central Brazil (Azmy et al. 2008; Geboy et al. 2013), though their age is somewhat enigmatic. Recent Re-Os geochronology on organic-rich shales below and above the diamictite yield an ages of ∼1.3 and ∼1.1 Ga, respectively, (Geboy et al. 2013), while the youngest detrital zircons in the diamictite itself are ∼0.9 Ga in age (Rodrigues et al. 2012). These deposits indicate that at least portions of the mid-Proterozoic were not entirely ice-free, but their deposition at relatively high paleolatitude (Tohver et al. 2006) renders their broader climatic implications somewhat enigmatic.

In any case, the close of the Proterozoic Eon (2.5–
0.541 Ga) bore witness to perhaps the most severe climate perturbations in Earth’s history, the Neoproterozoic “Snowball Earth” events (Hoffman et al. [1998]. Recent high-resolution Re-Os geochronology delineates two major glacial episodes, the protracted Sturtian glaciation (lasting between 717–660 Ma) and the shorter Marinoan glaciation (terminating at 635 Ma), with a relatively brief interglacial period lasting less than 25 million years (Rooney et al. 2015). While understanding the intensity, dynamics, and biogeochemical impacts of these glaciations remain areas of active research [extensively reviewed in Hoffman & Schrag (2002), Pierrehumbert et al. (2011), and [Hoffman et al. (2017)], it is clear that this period marks a dramatic perturbation to planetary climate and surface habitability.

The Phanerozoic Eon (e.g., the last 541 million years) has been marked by at least three large-scale ice ages, during the Ordovician (Delabroye & Vecoli 2010), the Perm-Carboniferous (Veevers & Powell 1987), and the late Cenozoic (Zachos et al. 2001). Of these, the late Paleozoic ice age was the most intense and long-lived (Montanez & Poulsen 2013). These events have been linked with faunal turnover and mass extinction (Raymond & Metz 2004), and in some cases reflect major milestones in the evolution of Earth’s biosphere such as the earliest colonization of the land surface by non-vascular plants (Lenton et al. 2012) and the extensive production and burial of organic matter by burgeoning terrestrial ecosystems (Feulner 2017). At the same time, the most recent half-billion years of Earth’s history shows evidence for significant transient perturbations to Earth’s carbon cycle and climate system on a wide range of timescales (Zachos et al. 2008) and often associated with dramatic changes to the diversity and abundance of macroscopic life (Erwin 1994, Payne et al. 2004). Nevertheless, despite large changes to carbon fluxes into and out of the ocean-atmosphere system Earth’s climate has consistently avoided the sort of catastrophic climate destabilization witnessed during the late Proterozoic.

In sum, Earth’s geologic record suggests that the establishment of a robust hydrosphere, with liquid water oceans and low-temperature aqueous alteration of exposed crust, occurred very shortly after Earth’s formation. In addition, surface temperatures have generally been stable and relatively warm for the vast majority of Earth’s history, despite long-term changes in solar insolation and dramatic changes to atmospheric composition (see below). However, this history also highlights the importance of internal feedbacks within the climate system in structuring planetary habitability on Earth (and thus likelihood of remote detection) over time. In particular, the “Snowball Earth” glaciations suggest that an Earth-like planet that spends its lifetime safely within the Habitable Zone of its host star can still undergo catastrophic climate destabilization, and that both the timing and duration of these events can be unpredictable (e.g., Rooney et al. 2015). This is in marked contrast to the limit-cycle climate instability predicted for Earth-like planets near the outer edge of the Habitable Zone (Haqq-Misra et al. 2016). In addition, the contrasting timescales of the Sturtian and the Marinoan glaciations imply a wide range of potential effects on the long-term maintenance of remotely detectable biosignatures, placing significant impetus on better understanding the large-scale biogeochemistry of “Snowball Earth” conditions (see below).

Geologic and geochemical data also provide a window into the dramatic evolution of Earth’s atmospheric chemistry, with implications for both the habitability of surface environments and the remote detectability of atmospheric biosignatures. We focus here on major changes to atmospheric gas species that are important for regulating global climate (e.g., CO₂, CH₄, N₂, and possibly H₂) and species that are potentially promising biosignature gases (e.g., O₂, O₃, CH₄, N₂O). Some species, most notably methane and organic hazes, serve dual roles as both arbiters of climate and potential biosignatures. We break the history of Earth’s atmospheric chemistry into four broad intervals: the Archean, the Paleoproterozoic, the mid-Proterozoic, and the Phanerozoic (Figure 11). However, we emphasize that these all represent extremely long periods of time, and that there is likely to be higher-order variability within each interval. Model-derived spectra of Earth at key evolutionary stages are shown in Figure 12 demonstrating how varying atmospheric compositions have led to dramatically different spectral appearances for our planet.

The apparently ubiquitous production and preservation of non-mass-dependent sulfur (NMD-S) isotope anomalies in marine sedimentary rocks of Archean age suggests extremely low atmospheric O₂ levels, constrained to an upper limit of ∼10⁻⁶ bar (Pavlov & Kasting 2002) but most likely below ∼10⁻⁸ bar (Claire et al. 2006, Zahnle et al. 2006). These anomalies also suggest high abundance of some reducing gas for effective production of S₈ in the atmosphere, with CH₄ as the most likely candidate (Zahnle et al. 2006). More recently, coherent time-dependent changes in quadruple-sulfur isotope systematics have been tied to the transient production and breakdown of atmospheric organic hazes during the late Archean (Zerkle et al. 2012, Izon et al. 2017). Given the surface CH₄ fluxes required to maintain persistent organic hazes, their presence may serve as an effective biosignature in reducing planetary atmospheres such as that of the Archean Earth (Arney et al. 2017). Isotopic evidence for a broadly reducing, low-O₂ ocean-atmosphere system is consistent with the presence of reduced detrital minerals in fluvial sediments (Rasmussen & Buck 1999), the loss of Fe and Mn from fossilized soil profiles (palaeosols) during initial weathering (Rye & Holland 1998), a pervasive lack of measurable cerium (Ce) anomalies in Fe-rich marine chemical sediments (Planavsky et al. 2010), and muted mass-dependent S isotope fractionation in marine sulfide minerals (Crowe et al. 2014).

The permanent disappearance of large NMD-S anomalies from the marine sedimentary rock record at ~2.3 billion years ago (e.g., Luo et al. 2016) marks the initial accumulation of O₂ in Earth’s atmosphere during the “Great Oxidation Event” (GOE), initially hypothesized on the basis of a general disappearance of detrital reduced minerals from flu-
Fig. 11.— Summary of theoretical and empirical constraints on the abundances of N$_2$, O$_2$, CO$_2$, and CH$_4$ in Earth’s atmosphere for the four major time periods discussed in the text (Archean, Paleoproterozoic, mid-Proterozoic, and Phanerozoic). Blue bars show reconstructions from models, while red bars show inferences based on empirical data. Also shown for the Archean are model-based estimates of prebiotic O$_2$ and CH$_4$ levels (grey bars). The ranges are meant to be inclusive, and some of the variability in a given time period should be considered to arise from time-dependent variability rather than uncertainty [e.g., Olson et al. (2018)]. Constraints are as described in Table 2.

Fig. 12.— Simulated spectra of Earth at key evolutionary stages. Colors indicate time period: Archean (orange), Paleoproterozoic (dark gray), mid-Proterozoic (light gray), and Phanerozoic (blue). Both hazy and haze-free Archean models are shown, and all models include fractional water cloud coverage. Key absorption features are indicated. Original sources for spectra are Arney et al. (2016) and Robinson et al. (2011).
vial sediments and the first appearance of extensive red beds at Earth’s surface (Holland 1984, 2002). More recently, geologic and geochemical evidence has led to the hypothesis of a protracted, but ultimately transient, period of ocean-atmosphere oxygenation following the GOE (reviewed in Lyons et al. 2014). In particular, marine sedimentary carbonate rocks record an extended interval of $^{13}C$ enrichment — the so-called ‘Lomagundi Event’ (Karhu & Holland 1996; Bekker 2001; Melezhik et al. 2007) — which implies a massive release of $O_2$ to the ocean-atmosphere system according to conventional models of Earth’s carbon cycle (Kump & Arthur 1999). This period also records the earliest extensive marine sulfate evaporite deposits (Schröder et al. 2008), expansion of mass-dependent $S$ isotope fractionations (Canfield 2005), a dramatic increase in the ratio of ferric to total iron in marine shales (Bekker & Holland 2012), significant enrichments of redox-sensitive metals in anoxic marine sediments (Canfield et al. 2013; Partin et al. 2013; Reinhard et al. 2013), and the first economic phosphorite deposits (Lepland et al. 2014).

A wide range of geochemical proxies point to a subsequent return to relatively low ocean-atmosphere oxygen levels during the mid-Proterozoic, between $\sim 1.8 - 0.8$ billion years ago. In particular, a disappearance of sulfate evaporites and large phosphorite deposits from the rock record (Schröder et al. 2008; Reinhard et al. 2017b), a drop in ferric to total iron ratios in marine shales (Bekker & Holland 2012), the $S$ isotope systematics of marine sulfide and sulfate phases (Planavsky et al. 2012; Scott et al. 2014), and the incomplete retention of Fe and Mn in lithified soil horizons during weathering (Zbinden et al. 1988) all point to a decrease in ocean-atmosphere oxygen levels following the Lomagundi Event. More recently, the stable chromium ($Cr$) isotope composition of marine sediments (Planavsky et al. 2014b; Cole et al. 2016), the iodine content (Hardisty et al. 2017) and rare Earth element systematics (Tang et al. 2016a) of shallow marine carbonate rocks, and the enrichments of redox-sensitive metals in anoxic marine shales (Partin et al. 2013; Reinhard et al. 2013; Sheen et al. 2018) have buttressed this view. The majority of geochemical observations are consistent with a background $pO_2$ value at or well below $\sim 10^{-2}$ bar (Lyons et al. 2014). However, NMD-$S$ anomalies do not return during this interval, suggesting that atmospheric $pO_2$ remained above $\sim 10^{-6}$ bar, atmospheric $pCH_4$ remained well below $\sim 10^2 - 10^3$ bar, or both (e.g., Zahnle et al. 2006). Precisely quantifying atmospheric $pO_2$ during this period remains a significant outstanding challenge, and Archean oxygen levels, perhaps paradoxically, are perhaps better constrained than those of the Proterozoic.

The late Proterozoic bore witness to significant changes in ocean-atmosphere redox, before, during, and after the “Snowball Earth” glaciation events (reviewed in Lyons et al. 2014). Indeed, there is some evidence for a shift in ocean-atmosphere redox immediately preceding the first low-latitude glaciation (Planavsky et al. 2014b; Thomson et al. 2015), implicating time-dependent changes to Earth’s oxygen cycle as a potentially important component of climate destabilization in both the Paleoproterozoic and Neoproterozoic. The ultimate result of these upheavals appears to have been an oxygenation of the ocean-atmosphere system to a degree approaching that of the modern Earth. For most of Phanerzoic time ($541$ million years ago to the present), atmospheric $pO_2$ appears to have remained within the “fire window” of between $\sim 0.15 - 0.35$ bar (Belcher & McElwain 2008; Glasspool & Scott 2010), although atmospheric $pO_2$ during the Paleozoic prior to the rise of land plants is somewhat poorly constrained and may have been well below $0.1$ bar (e.g., Bergman 2004; Lenton et al. 2018). However, essentially all available geologic, geochemical, and biological observations are consistent with a well-oxygenated ocean-atmosphere system for the last $500-600$ million years (Lyons et al. 2014).

Estimates of atmospheric $CO_2$ during the Archean and Proterozoic cover a very wide range. Mineral assemblages in authigenic rinds on riverine gravels provide a lower limit on atmospheric $pCO_2$ of roughly $10^{-3}$ bar at $3.2$ Ga (Hessler et al. 2004). However, these calculations require an assumption of thermodynamic equilibrium during transport and alteration, which may not be strictly valid, and are in any case consistent with $CO_2$ levels an order of magnitude or more higher than this at the range of surface temperatures implied by the isotopic constraints discussed above. Similarly, secondary mineral assemblages in ancient soil horizons that formed between $\sim 2.7 - 2.5$ Ga have been interpreted to indicate $pCO_2$ values between $\sim 10^{-3} - 10^{-2}$ bar during the late Archean (Rye et al. 1995; Sheldon 2006). A more recent model suggests much higher Archean $pCO_2$, up to or exceeding $\sim 10^{-1}$ bar (Kanzaki & Murakami 2015), though estimates according to this method can vary over many orders of magnitude at any given time due to uncertainties in assumed soil formation timescales. Archean atmospheric $pCO_2$ is thus not very well-constrained, though all existing data are consistent with values that were elevated above those of the modern Earth, perhaps by $2-3$ orders of magnitude.

The same approach applied to paleosols formed between $2.5-1.8$ Ga yields a somewhat more consistent picture, with estimates from both of the most recent $pCO_2$ reconstruction techniques yielding values on the order of $\sim 10^{-3} - 10^{-2}$ bar (Sheldon 2006; Kanzaki & Murakami 2015), although the estimates of Kanzaki & Murakami (2015) for similarly aged paleosols are in general a factor of $2-5$ higher than those of Sheldon (2006). There is only one well-studied paleosol near the Archean-Proterozoic boundary, making it difficult to establish with confidence whether significant changes in atmospheric $pCO_2$ occurred transiting the Archean-Proterozoic boundary. Moving into the mid-Proterozoic, reconstructions based on paleosols (Sheldon 2013), the carbon isotope compositions of individual microfossils (Kaufman & Xiao 2003), and calcification patterns of cyanobacterial fossils (Kah & Riding 2007) are all broadly consistent with atmospheric $pCO_2$ values on the order of $\sim 3 \cdot 10^{-3}$ bar, though individual estimates vary between values roughly equivalent to the modern Earth to
high values of $\sim 10^{-2}$ bar. Broadly, these observations tend to suggest a drop in atmospheric CO$_2$ levels between the Paleoproterozoic and the mid-Proterozoic.

A much wider range of potential $p$CO$_2$ proxies exists for the Phanerozoic, including a higher-resolution paleosol record, the carbon isotope compositions of a wide range of plant and phytoplankton taxa, the density of stomata on fossilized leaves, and geochemical proxies related to the pH-dependent speciation of boron in seawater (reviewed in Royer 2014). Although these approaches are all undergoing continual refinement, they generally point to a range for atmospheric $p$CO$_2$ during the Phanerozoic between roughly $\sim 10^{-4}$--$10^{-3}$ bar, with coherent secular shifts associated with major biospheric innovations and changes in climate (see above). An important caveat to this record is that constraints become very patchy during early Paleozoic time (e.g., prior to round 400 million years ago), a period during which Earth system models indicate atmospheric CO$_2$ levels were higher than at any other time during the Phanerozoic (Berner & Kothava, 2001). However, most recent inversions suggest that atmospheric $p$CO$_2$ has never been significantly above $\sim 5 \times 10^{-3}$ bar during the last 500 million years (Royer et al. 2014; Lenton et al. 2018).

A current frontier in reconstructing the evolution of Earth’s atmosphere is developing constraints on atmospheric pressure, as linked most directly with changes in atmospheric N$_2$ abundance. As discussed below, mass balance calculations suggest that N$_2$ levels may have varied significantly from the present level of $\sim 0.8$ bar, with potentially non-trivial impacts on climate (Goldblatt et al. 2009). Recent approaches toward reconstructing overall atmospheric pressure have included estimating air density based on the diameter of fossilized raindrop imprints in a 2.7-billion-year-old tuff from the Ventersdorp Supergroup, South Africa (Som et al. 2012), and estimating total barometric pressure via the size distribution of vesicles in a basalt flow from roughly the same age preserved in the Pilbara Craton, Australia (Som et al. 2016). The technique based on raindrop imprints implies that atmospheric pressure at 2.7 Ga was between $\sim 0.5$--2.0 bar, though placing an upper limit with this method is difficult due to potential variability in rain rates (Kavanagh & Goldblatt 2015). The vesicular basalt approach provides a much more stringent upper limit of around 0.5 bar. Marty et al. (2013) attempted to estimate $p$N$_2$ directly by analyzing the isotopic composition of nitrogen and argon in 3.5 and 3.0 billion-year-old fluids trapped in hydrothermal quartz from the Pilbara craton, Australia, deriving mixing arrays between end-member hydrothermal fluids of variable composition with a single end-member for air-saturated Archean seawater. Their analysis indicates that $p$N$_2$ was not significantly above $\sim 0.5$--1.0 bar during the early Archean. Nishizawa et al. (2007) provide a $p$N$_2$ estimate of $\sim 3$ bar from fluid inclusions in the same unit, but the N$_2$/$^{38}$Ar values from their samples indicate that their data do not capture the low-N$_2$ end-member analyzed by Marty et al. (2013) and thus likely overestimate ambient $p$N$_2$. In any case, uncertainties in all current approaches and the fragmentary nature of the archives required for their application allow for atmospheric N$_2$ abundance to vary by a factor of two or more above/below modern, rendering the potential climate impacts of N$_2$ somewhat enigmatic at present but important to consider (see below).

5.2. Models

Standard stellar evolution models (Gough 1981) predict that the Sun was 20--30% less luminous than it is today during the Hadean and Archean. The observation of a generally clement or even warm climate during the Hadean and Archean (see above) thus implies that the composition of Earth’s atmosphere was very different from that of the modern. Indeed, as discussed above there is persuasive geological and geochemical evidence that the composition of Earth’s atmosphere was very different during the Hadean, Archean, and Proterozoic.

The most prominent solutions to this problem invoke a larger inventory of greenhouse gases in Earth’s early atmosphere. Sagan & Mullen (1972) explored a reducing NH$_3$--CH$_4$--H$_2$--H$_2$O--CO$_2$ greenhouse, with a dominant role for NH$_3$. However, the rapid photolysis of NH$_3$ in the upper atmosphere would have required a very large source at Earth’s surface, and would have in turn resulted in the production of rather extreme amounts of N$_2$ on geologically rapid timescales (Kuhl & Atreya 1979). More recently, it has been suggested that the photolysis of NH$_3$ in the upper atmosphere may have been mitigated somewhat by absorption of UV photons by a fractal organic haze (Wolf & Toon 2010), an idea that warrants additional scrutiny of the relative altitudes of peak NH$_3$ photolysis and haze absorption in future work (Wolf & Toon 2010). In any case, subsequent work has tended to focus on CH$_4$--CO$_2$--H$_2$O greenhouses and, more recently, the possible radiative effects of high H$_2$.

Models predict that the abundance of CH$_4$ in Earth’s prebiotic atmosphere would have been low (Kasting 2005; Emanuel & Ague 2007). Atmospheric CO$_2$ during this period is poorly constrained, but a recent inversion using a geologic carbon cycle model yields a median $p$CO$_2$ estimate of 0.3 bar, with a 95% confidence interval of 0.03--1.0 bar (Krissansen-Totton et al. 2018). Greenhouses dominated by H$_2$O and CO$_2$ with $p$CO$_2$ values on the lower end of this range would be unlikely to exhibit cloud top surface temperatures under Hadean or early Archean solar luminosity, but both 1-D radiative-convective and 3-D global climate models predict that values at the upper end of this range would result in surface temperatures well above freezing under early and late Archean luminosity (Kasting & Ackerman 1986; Kasting 1987; Charnay et al. 2013; Wolf & Toon 2013, 2014). Significant additional warming may have been provided by collision-induced absorption by H$_2$--N$_2$ under plausible prebiotic conditions (Wordsworth & Pierrehumbert 2013), though the strength of this would have depended
| Period          | Case            | $p_{N_2}$ [bar] | $p_{O_2}$ [bar] | $p_{CO_2}$ [bar] | $p_{CH_4}$ [bar] | Source(s)              |
|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|------------------------|
| prebiotic       | model high      | $3 \times 10^{-12}$       | –               | $3 \times 10^{-5}$       | 1.2             |                        |
|                 | model low       | $3 \times 10^{-14}$       | –               | $5 \times 10^{-6}$       | 1.3             |                        |
|                 | data high       | –                   | –               | –                | –               |                        |
|                 | data high       | –                   | –               | –                | –               |                        |
| Archean         | model high      | 2.3               | $2 \times 10^{-8}$ | 1.0 $\times 10^0$ | 5.0 $\times 10^{-1}$ | 4-12                   |
|                 | model low       | 0.3               | $1 \times 10^{-10}$ | 3.0 $\times 10^{-3}$ | 1.0 $\times 10^{-4}$ | 5-13                   |
|                 | data high       | 1.1               | $2 \times 10^{-6}$ | 7.0 $\times 10^{-1}$ | 1.4 $\times 10^{-1}$ | 14-16                  |
|                 | data low        | 0.1               | –               | 7.0 $\times 10^{-3}$ | 1.0 $\times 10^{-4}$ | 7,17-19                |
| Paleoproterozoic| model high      | 0.8               | $3 \times 10^{-1}$ | 1.0 $\times 10^{-1}$ | 5.0 $\times 10^{-6}$ | 8-10,13,20,21         |
|                 | model low       | 0.3               | $2 \times 10^{-2}$ | 2.0 $\times 10^{-3}$ | 7.0 $\times 10^{-7}$ | 8-10,13,21             |
|                 | data high       | –                 | 3.0 $\times 10^{-1}$ | 1.5 $\times 10^{-1}$ | –               | 16,22                  |
|                 | data low        | –                 | 1.0 $\times 10^{-3}$ | 7.0 $\times 10^{-3}$ | –               | 18,22,23               |
| mid-Proterozoic | model high      | 0.8               | $2 \times 10^{-2}$ | 8.0 $\times 10^{-2}$ | 3.0 $\times 10^{-5}$ | 8-10,13,24-26         |
|                 | model low       | 0.4               | $6 \times 10^{-4}$ | 6.0 $\times 10^{-4}$ | 5.0 $\times 10^{-6}$ | 8-10,13,25,27         |
|                 | data high       | –                 | 8.0 $\times 10^{-3}$ | 5.5 $\times 10^{-2}$ | –               | 23,28,29               |
|                 | data low        | –                 | 2.0 $\times 10^{-5}$ | 3.0 $\times 10^{-4}$ | –               | 30,31                  |
| Phanerozoic     | model high      | 0.9               | $3 \times 10^{-1}$ | 5.5 $\times 10^{-3}$ | 1.0 $\times 10^{-5}$ | 32-36                  |
|                 | model low       | 0.7               | $4 \times 10^{-2}$ | 2.8 $\times 10^{-4}$ | 1.0 $\times 10^{-7}$ | 32-36                  |
|                 | data high       | 0.8               | $1.5 \times 10^{-1}$ | 2.8 $\times 10^{-3}$ | 8.0 $\times 10^{-7}$ | 37-39                  |
|                 | data low        | 0.8               | $3 \times 10^{-1}$ | 1.9 $\times 10^{-4}$ | 3.5 $\times 10^{-7}$ | 39-41                  |

References: — 1 Haqq-Misra et al. (2011), 2 Emmanuel & Aigue (2007), 3 Tian et al. (2011), 4 Goldblatt et al. (2009), 5 Claire et al. (2006), 6 Zahnle et al. (2006), 7 Kurzweil et al. (2013), 8 Kasting (1987), 9 Haley & Bachan (2017), 10 Krissansen-Totton et al. (2018), 11 Kharecha et al. (2005), 12 Ozaki et al. (2018), 13 Stueken et al. (2016), 14 Marty et al. (2013), 15 Pavlov & Kasting (2002), 16 Kanzaki & Murakami (2015), 17 Som et al. (2016), 18 Sheldon (2006), 19 Driese et al. (2011), 20 Bachan & Kump (2015), 21 Harada et al. (2015), 22 Bekker & Holland (2012), 23 Rye & Holland (1998), 24 Laakso & Schrag (2017), 25 Catling et al. (2007), 26 Zhao et al. (2018), 27 Olson et al. (2016), 28 Kaufman & Xiao (2003), 29 Kah & Riding (2007), 30 Planavsky et al. (2014b), 31 Sheldon (2013), 32 Berner (2006), 33 Royer (2014), 34 Lentz et al. (2018), 35 Bartdorff et al. (2008), 36 Beerling et al. (2009), 37 Belcher & McElwain (2008), 38 Royer (2014), 39 Wolff & Spahni (2007), 40 Glasspool & Scott (2010), 41 Galbraith & Eggleston (2017)

† Assumes $p_{CO_2} = 2$ bar.
‡ Assumes $p_{CO_2} = 0.02$ bar.
§ Assuming data high $p_{CO_2}$ and $CH_4/CO_2 = 0.2$.
¶ Assuming predominantly $N_2$ atmosphere.
 đám Assuming $p_{O_2} = 10^{-3}$ bar, $[SO_4^{2-}] = 500$ mol/kg.

Note.—All values are approximate. See primary references for assumptions and caveats not noted here.
strongly on atmospheric H$_2$ and N$_2$ abundance, both of which are poorly constrained for the prebiotic atmosphere.

The emergence of a biosphere on Earth would have had a significant impact on atmospheric chemistry and climate. In particular, primitive microbial metabolisms such as methanogenesis, acetogenesis, and anoxygenic photosynthesis would have dramatically increased fluxes of CH$_4$ to Earth’s atmosphere. The implications of this for climate are twofold. First, CH$_4$ is an important greenhouse gas in its own right, providing another means toward offsetting decreased solar luminosity that would be particularly effective in a reducing atmosphere (Pavlov et al. 2000; Haqq-Misra et al. 2008). Second, the potential for large biogenic CH$_4$ fluxes introduces an additional feedback on climate via formation of an organic haze in the atmosphere (Zahnle 1986; Pavlov et al. 2001; Haqq-Misra et al. 2008). Photochemical models (Haqq-Misra et al. 2008; Zerkle et al. 2012; Arney et al. 2016) and laboratory experiments (DeWitt et al. 2009; Trainer et al. 2004, 2006) predict that once the CH$_4$/CO$_2$ ratio of the atmosphere increases beyond $\sim$0.1 hydrocarbon aerosols will begin to form, with optical thicknesses at visible and UV wavelengths that increase rapidly at CH$_4$/CO$_2$ values between $\sim$0.1–1 (Domagal-Goldman et al. 2008; Haqq-Misra et al. 2008). The formation of these hazes can lead to significant cooling, even at relatively low CH$_4$/CO$_2$ values of $\sim$0.1 – 0.2 (Arney et al. 2016). These combined effects would have been important in regulating climate and surface temperature during both the Archean and Proterozoic.

Taken together, models suggest that element or even warm surface temperatures could have been maintained on the prebiotic Earth by a CO$_2$–H$_2$O greenhouse, potentially supplemented by collision- and pressure-induced warming at greater H$_2$ and N$_2$ abundance (Kasting 1987; Goldblatt et al. 2009; Wordsworth & Pierrehumbert 2013; Krissansen-Totton et al. 2018). Following the emergence of a surface biosphere, 1-D radiative-convective and 3-D global climate models, coupled ecosystem-biogeochemistry models, and the geologic record are all consistent in suggesting that surface temperatures well above freezing could have been maintained throughout the Archean with a CH$_4$–H$_2$O–CO$_2$ greenhouse and optically thin haze (Haqq-Misra et al. 2008; Zerkle et al. 2012; Charnay et al. 2013; Wolf & Toon 2013; Izon et al. 2017; Ozaki et al. 2018; Krissansen-Totton et al. 2018). However, additional factors beyond changes to the atmospheric greenhouse may also have been important, and together could lead to significant warming. For example, changes to the average size of cloud droplets attendant to fewer cloud condensation nuclei (CCN) in the early atmosphere would have resulted in more effective rainout and fewer low clouds at low-mid latitudes (Charnay et al. 2013; Wolf & Toon 2014) and possibly thinner clouds overall (Goldblatt & Zahnle 2011; Charnay et al. 2013), both of which would have resulted in significant warming. Maintaining a habitable climate during Earth’s earliest history is thus not particularly challenging, as all of these factors were potentially in play. However, achieving the highest estimates of Archean temperature with plausible atmospheric CO$_2$ and CH$_4$ levels remains a challenge.

Existing models of the long-term carbon cycle and surface temperature are broadly consistent with geochemical evidence for elevated pCO$_2$ during the Paleoproterozoic (e.g., Haley & Bachan 2017; Krissansen-Totton et al. 2018). However, the initial rise of atmospheric O$_2$ at $\sim$2.3 Ga would have destabilized the Archean atmospheric greenhouse (Claire et al. 2006; Zahnle et al. 2006; Haqq-Misra et al. 2008), potentially leading to dramatic effects on climate during the earliest Paleoproterozoic (see above). Understanding climate dynamics and coherently modeling climate and biogeochemistry during and after the GOE and through the Lomagundi Event remain outstanding challenges, but existing models suggest large changes to Earth surface temperature, the global carbon cycle, and atmospheric composition (Claire et al. 2006; Harada et al. 2015).

There is limited geologic evidence for glacial conditions during the mid-Proterozoic (see above), despite a solar luminosity of $\sim$10% relative to the modern Earth (Gough 1981). Recent coupled 3-D climate modeling indicates that global glaciation should occur under these conditions if atmospheric pCO$_2$ drops to around $10^{-3}$ bar (Fiorella & Sheldon 2017), and extensive glaciation at high and middle latitudes should occur even above pCO$_2$ values an order of magnitude higher than the modern Earth unless surface temperatures are buffered by some other greenhouse gas. Nitrous oxide (N$_2$O) would be unlikely to provide the requisite radiative forcing, given the relatively low atmospheric pO$_2$ during the mid-Proterozoic and the large biological nitrogen fluxes required (Roberson et al. 2011). Methane (CH$_4$) is another candidate, but 3-D models of ocean biogeochemistry (Olson et al. 2016) interpreted in light of 3-D climate modeling (Fiorella & Sheldon 2017) indicate that an ocean-only CH$_4$ cycle would have been unable to maintain an ice-free climate during the mid-Proterozoic. One possible solution to this would be a significant microbial CH$_4$ flux from terrestrial ecosystems (Zhao et al. 2018). Alternatively, atmospheric pCO$_2$ may have been somewhat higher than geochemical proxies suggest (Krissansen-Totton et al. 2018), or other changes to factors like cloud droplet radius or surface albedo may have contributed to stabilizing relatively warm temperatures (Fiorella & Sheldon 2017). Lastly, at least periods of the mid-Proterozoic may not have been entirely ice-free (see above). In any case, climate models, geochemical proxies, and models of marine/terrestrial biogeochemistry yield a picture of a relatively weak H$_2$O–CO$_2$ greenhouse buffered by CH$_4$ levels on the order of $\sim$10$^{-5}$ bar or slightly less, depending on the importance of terrestrial CH$_4$ cycling.

The dynamics of Earth’s climate during the intense ice ages of the late Proterozoic are much more well-studied than their Paleoproterozoic counterparts, and the reader is here referred to two recent comprehensive reviews on the subject (Pierrehumbert et al. 2011; Hoffman et al. 2017). However, the biogeochemical dynamics associated with these perturbations are more poorly understood, particu-
larly with regard to Earth’s O₂ and CH₄ cycles, and this represents an important topic of future work. For example, although impacts to the Earth’s "oxidized" carbon cycle have been explored in a range of models (Le Hir et al. 2008b,a; Mills et al. 2011), it remains unclear what role Earth’s CH₄ cycle may have played in the inception or recovery from low-latitude glaciation, if any (Schrag et al. 2002; Pierrehumbert et al. 2011; Olson et al. 2016). In any case, the Sturtian glacial episode in particular is estimated to have lasted for roughly 50 million years (Rooney et al. 2014), with the attendant impacts on atmospheric biosignatures and the remote detectability of any surviving biosphere almost completely unknown.

The Phanerzoic climate system, though perhaps relatively stable in the scope of Earth’s entire history, has been extremely dynamic, with at least three major ice ages and intervening periods of relatively warm, largely ice-free conditions (see above). Though they differ in their tectonic interregnums and the remote detectability of any surviving biosphere...
The GOE effectively represented a shift in the trace redox gas in Earth’s atmosphere from O$_2$ to CH$_4$. Although models do indeed predict an initial drop in atmospheric pCH$_4$ during the GOE, the ultimate establishment of a substantial stratospheric O$_3$ layer attendant to rising ground-level pO$_2$ is predicted to shield CH$_4$ from destruction in the troposphere. This allows atmospheric CH$_4$ levels to rebound in photochemical models to $\sim 10^{-4}$ bar in the Paleoproterozoic (Claire et al. 2006; Goldblatt et al. 2006). However, models of that include ocean biogeochemistry and microbial consumption of CH$_4$ with O$_2$ and SO$_4^{2-}$ generally result in lower steady-state atmospheric pCH$_4$ following the GOE, typically on the order of $\sim 10^{-5}$ bar or lower (Catling et al. 2007; Daines & Lenton 2016; Olson et al. 2016). These results depend strongly on assumed atmospheric pO$_2$ and the ocean reservoir of SO$_4^{2-}$ (Olson et al. 2016), and are not currently equipped to deal with the potentially important impacts of a terrestrial biosphere (Zhao et al. 2018). A full exploration of this problem will require coupled, open-system models of photochemistry and ocean/terrestrial microbial metabolism. Nevertheless, the abundance of CH$_4$ in Earth’s atmosphere following the GOE was likely significantly lower than that of the Archean Earth, and in particular would have been orders of magnitude below that of the earliest Archean and Hadean Earth prior to the evolution of oxygenic photosynthesis (Kharecha et al. 2005; Ozaki et al. 2018). Unfortunately, a quantitative geologic or geochemical indicator of atmospheric pCH$_4$ at levels below those of the Archean has not yet been developed, and atmospheric pCH$_4$ is very difficult to track empirically on the post-Archean Earth.

Although the differences in mean state before and after the GOE are relatively well understood, the dynamics of climate and atmospheric chemistry in the immediate aftermath of the GOE are not. Some models predict that this change to Earth’s surface redox balance would have had significant climate impacts (Claire et al. 2006; Haqq-Misra et al. 2008), one result of which may have been an ultimately transient but quantitatively dramatic elevation in atmospheric pO$_2$ (Harada et al. 2015). This scenario would be consistent with emerging geochemical evidence for elevated atmospheric O$_2$ (and thus O$_3$), possibly for 100-million-year timescales, during the Paleoproterozoic (see above). The protracted, but ultimately transient, rise in atmospheric pO$_2$ implies a significant drop in atmospheric CH$_4$ levels (Harada et al. 2015), and potentially a substantial drop in atmospheric pCO$_2$ unless buffered by a sedimentary rock cycle very different from that of the modern Earth (e.g., Bachan & Kump 2015). In any case, long-term (e.g. mean state) carbon cycle and climate models that are entirely uncoupled or only implicitly coupled to the O$_2$ cycle are broadly consistent with the current geochemical constraints for atmospheric pCO$_2$ during the Paleoproterozoic discussed above (Sleep & Zahnle 2001; Haley & Bachan 2017; Krissansen-Totton et al. 2018).

The initial accumulation of O$_2$ in Earth’s atmosphere appears to have been followed by a subsequent return to
relatively low ocean-atmosphere oxygen levels during the mid-Proterozoic (between \(1.8 \times 0.8\) billion years ago). However, the absence of non-mass-dependent S isotope fractionations in marine sediments and the apparent absence of reduced detrital minerals in fluvial settings indicate atmospheric \(pO_2\) remained above \(10^{-6}\) bar. On long timescales, the modern atmospheric \(O_2\) level is maintained dynamically by the balance between net \(O_2\) sources (principally the burial of organic carbon and reduced sulfur into the Earth’s upper crust) and net \(O_2\) sinks (largely the subsequent exhumation and oxidative weathering of organic carbon and reduced sulfur, along with reactions between \(O_2\) and reduced metamorphic and volcanic gases). However, there are strong nonlinearities in the scaling relationships between these fluxes and the amount of \(O_2\) in the atmosphere. In addition, the major sink fluxes on the modern Earth decrease in magnitude as atmospheric \(pO_2\) drops, while the major source fluxes increase. As a result, not all atmospheric \(pO_2\) values are equally stable, and understanding the internal processes and feedbacks capable of maintaining atmospheric \(O_2\) levels above those characteristic of the Archean but well below those of the modern Earth remains an outstanding question (Lyons et al. 2014; Daines et al. 2017).

As discussed above, models of mid-Proterozoic climate and biogeochemistry suggest a relatively weak \(H_2O-CO_2\) greenhouse buffered by modest \(CH_4\) levels (Figure 11, Table 2). In particular, long-term carbon cycle models suggest atmospheric \(pCO_2\) of around \(10^{-3}-10^{-2}\) bar during the mid-Proterozoic, while models of marine/terrestrial biogeochemistry suggest atmospheric \(pCH_4\) values on the order of \(10^{-6}-10^{-5}\) bar, together consistent with most geochemical constraints and a largely ice-free climate state (Fiorella & Sheldon 2017). That said, some paleosol reconstructions approach roughly modern \(pCO_2\) values (e.g., Sheldon 2013), which is difficult to reconcile with evidence for a largely ice-free Earth surface for most of the mid-Proterozoic unless Earth’s greenhouse was impacted strongly by fluxes of \(CH_4\) from a terrestrial microbial biosphere (Zhao et al. 2018). A full picture of Earth’s mid-Proterozoic atmosphere awaits a comprehensive model that couples open-system carbon cycling with a balanced redox budget and dynamic \(O_2-CH_4\) cycle, but existing data and models are consistent with this period of Earth’s history representing a potential “false negative” for conventional biosignature techniques (e.g., Reinhard et al. 2017).

Though there is accumulating geologic and geochemical evidence that the extreme low-latitude glaciations of the late Proterozoic were associated with significant changes to ocean-atmosphere redox and atmospheric chemistry (Hoffman et al. 1998; Canfield et al. 2007; Sahoo et al. 2012; Cox et al. 2013; Planavsky et al. 2014b; Thomson et al. 2015; Hoffman et al. 2017), the relative timing and mechanistic links remain somewhat obscure. Low-order biogeochemical modeling indicates that low-latitude glacial episodes can readily drive a secular transition from low- to high-oxygen steady states at sufficiently high \(pCO_2\) thresholds for deglaciation (Laakso & Schrag 2017). For example, a deglaciation threshold of \(pCO_2\) of \(0.1\) bar is sufficient to drive a permanent transition in atmospheric \(pCO_2\) from \(\sim 10^{-3}\) to \(10^{-1}\) bar (Laakso & Schrag 2017) during deglaciation. Glacial \(CO_2\) levels of this order are readily achievable even in models that allow for efficient ocean-atmosphere gas exchange and seafloor weathering (e.g., Le Hir et al. 2008b). Efforts to better understand the temporal polarity and mechanistic details linking climate destabilization and nonlinear changes to atmospheric chemistry during both the Paleoproterozoic and Neoproterozoic represent an important avenue of future work. Nevertheless, significant changes in the redox state of Earth’s ocean-atmosphere system are strongly implicated as having been both cause and consequence of sporadic perturbations to Earth’s habitability.

Biogeochemical models are generally consistent in suggesting a high-\(O_2\), low-\(CH_4\), and moderate \(CO_2\) atmosphere throughout the Phanerozoic. Both atmospheric \(O_2\) and atmospheric \(CO_2\) have been controlled by the combined effects of roughly modern solar luminosity, time-dependent variability in volcanic degassing, rock uplift, changes to the major ion chemistry of seawater, and the emergence and expansion of terrestrial ecosystems (Berner 1991; 2006; Royer et al. 2014; Lenton et al. 2018). Long-term atmospheric \(CH_4\) levels have been controlled largely by the evolutionary and climate dynamics controlling biogenic \(CH_4\) fluxes from terrestrial ecosystems (Bartford et al. 2008). Despite some discrepancies between different models in estimates of atmospheric \(pO_2\) during the earliest Paleozoic, most models indicate ranges for atmospheric \(pO_2\), \(pCO_2\), and \(pCH_4\) between \(0.1 - 0.3\) bar, \(10^{-4} - 10^{-3}\) bar, and \(10^{-7} - 10^{-5}\) bar, respectively, all of which are consistent with existing geologic and geochemical constraints (Figure 11, Table 2). Similar models for atmospheric \(pN_2\) through time suggest values close to that of the modern Earth for most of the Phanerozoic, though direct geologic constraints on this are lacking for all but the most recent periods of Earth’s history.

6. DECIPHERING EXO-EARTH OBSERVATIONS

The observations and models discussed in previous sections provide insights into remote sensing approaches to understanding distant habitable worlds. At wavelengths spanning the ultraviolet through the infrared, and for both broadband photometry and spectroscopy across a range of resolutions, data (or simulated data) for the distant Earth contain a great deal of information about the planetary environment. The sections below discuss the information content of photometry and spectroscopy in reflected light as well as observations at thermal infrared wavelengths. Additional information can be found in reviews by Meadows (2008), Kaltenegger et al. (2010), Kaltenegger et al. (2012), and Robinson (2018).
6.1. Visible Photometry

Single-instance broadband photometry of a distant Earth-like world provides limited information about the planetary environment. As Earthshine observations have shown (Qui et al. 2003; Pallé et al. 2003), photometric data could constrain planetary albedo — which is central to an understanding of planetary energy balance — as long as planetary size and phase are known. (If the planetary radius is unknown, the planetary reflectivity and size are degenerate.) Additionally, broad absorption features can be detected using photometric observations (e.g., as was the case for the 950 nm H$_2$O band in EPOXI observations; Livengood et al. 2011), although constraining atmospheric abundances from low-resolution observations is extremely challenging (Lupu et al. 2016). Planetary color derived from broadband observations has been suggested as a means of identifying exo-Earth candidates (Traub 2003), although more recent results indicate that this approach is problematic for Earth at any geological phase other than the Phanerozoic (Krissansen-Totton et al. 2016) and could be confused by planetary phase effects as well as our lack of knowledge on realistic colors of temperate or cool exoplanets.

Disk-integrated photometric observations that resolve the rotation of an exo-Earth yield much more powerful diagnostics than single-instance photometry (Figure 13). When acquired over multiple days (rotations), the rotation rate of the planet can be determined from diurnal variability (which is typically 10–20%; Ford et al. 2001; Livengood et al. 2011), even in the presence of evolving weather patterns (Pallé et al. 2008; Oakley & Cash 2009). Once the rotation rate of an exo-Earth is known, the correspondence between time and sub-observer longitude enables longitudinally-resolved mapping (Cowan et al. 2009; Fujii et al. 2010; Kawahara & Fuji 2010; Fujii et al. 2011; Fujii & Kawahara 2012; Cowan & Strait 2013), although degeneracies can occur in mapping approaches (Fujii et al. 2017). Additionally, studying photometric variability inside absorption bands of well-mixed gases (e.g., O$_2$) as compared to variability inside bands of other species (e.g., H$_2$O) can reveal condensation processes in planetary atmospheres (Fujii et al. 2013). Depending on the optical thickness of a potential haze in the atmosphere of the Archean Earth (Arney et al. 2016), it might be necessary to push photometric observations to red or near-infrared wavelengths to have surface and near-surface sensitivity in lightcurves. Photometric exo-Earth observations resolved at both rotational and orbital timescales could reveal additional information about the planetary surface. Due to the obliquity of the planetary rotational axis, and depending on orbital inclination, maps resolved in latitude and longitude could be produce from high-quality data (Fujii & Kawahara 2012; Cowan et al. 2013; Schwartz et al. 2016). Even in the absence of rotationally-resolved photometry, surface oceans — whose presence directly confirms the habitability of an exoplanet — could be detected via the effect of specular reflectance on a planetary phase curve (McCullough 2006; Williams & Gaidos 2008), especially at red and near-infrared wavelengths where observations at large phase angles are less strongly impacted by Rayleigh scattering and, thus, have better surface sensitivity (Trafton et al. 2010; Zugger et al. 2011) (observations at these wavelengths would also be less influenced by any hazes on the Archean Earth; Arney et al. 2016). Additionally, scattering at the Brewster angle will maximize the polarization signature from an exo-ocean and could be detected in the polarization phase curve of an exo-Earth (McCullough 2006; Stas08 [Williams & Gaidos 2008; Zugger et al. 2010]). Finally, polarization and reflectance phase curves for Earth-like exoplanets can also reveal cloud properties through scattering effects (Bailey 2007; Karalidi et al. 2011, 2012; Karalidi & Stam 2012).

6.2. Visible Spectroscopy

Spectroscopic observations in reflected light provide powerful information about the atmospheric and surface environment of the Pale Blue Dot at any stage in its evolution. In addition to the insights offered from photometry (as spectra can always be degraded to lower resolution), observations at even moderate spectral resolution enable the detection of trace atmospheric gases. For example, Drossart et al. (1993) used near-infrared Galileo data to constrain the abundances of CO$_2$, H$_2$O, CO, O$_3$, CH$_4$, and N$_2$O in the atmosphere of Earth. While the Drossart et al. (1993) study used spatially-resolved observations, the same gaseous absorption features appear in the disk-integrated EPOXI dataset (Livengood et al. 2011).

Beyond trace gas detection, spectroscopic reflected-light observations can also constrain atmospheric pressure — a key determinant of habitability — through Rayleigh scattering effects, broadening of gas absorption lines and bands, and through collision-induced absorption and dimer features. Pressure-induced absorption features due to O$_2$ and N$_2$ occur throughout the near-infrared (and into the mid-infrared; Misra et al. 2014a; Schwieterman et al. 2015a). Of course, interpretation of Rayleigh scattering features and pressure-broadened absorption bands is not straightforward. The former depends on surface gravity and atmospheric mean molecular weight and can be masked by haze absorption at blue wavelengths, while the latter is impacted by the composition of the background atmosphere (e.g., Hedges & Madhusudhan 2016).

Feng et al. (2018) investigated retrievals of planetary and atmospheric properties for the modern Pale Blue Dot from visible-wavelength (0.4–1.0 μm) spectroscopy at a variety of spectral resolutions and signal-to-noise ratios. Here, firm constraints on key gas mixing ratios (for H$_2$O, O$_3$, and O$_2$), total atmospheric pressure, and planetary radius could be achieved with simulated Earth observations at a V-band signal-to-noise ratio of 20 and spectral resolution ($\lambda/\Delta\lambda$)
of 140. Thus, observations of this quality for modern exo-Earth twin could be sufficient to indicate that the planet is either super-Earth or Earth-sized and that O$_2$ is a major atmospheric constituent, thereby strongly indicating that the planet is inhabited.

6.3. Thermal Infrared Observations

Owing to the great technical challenges posed by techniques for observing Earth-like planets around Sun-like stars at long wavelengths, relatively little attention has been focused on understanding disk-integrated observations of our planet at thermal infrared wavelengths. Nevertheless, infrared spectra of the distant Earth — even at relatively low spectral resolution — provide a great deal of information about the atmospheric and surface environment. Most fundamentally, and unlike reflected-light data, infrared observations can directly constrain the radius of an exo-Earth. The flux received from a true blackbody depends on its temperature, size, and distance. If the distance to a target star is known, and with the temperature constrained via Wien’s displacement law, the size of a planet can then be determined from low-resolution thermal infrared observations. Of course, Earth is not emit like a true blackbody, which would introduce some uncertainty into a fitted planetary radius.

Beyond planetary size, infrared gas absorption features, by definition, reveal the key greenhouse gases of a planetary atmosphere. For Earth, observations (Figure 17) plainly reveal signatures of CO$_2$, H$_2$O, O$_3$, CH$_4$, and N$_2$O (Christensen & Pearl [1997], Hearty et al. [2009]). As molecular absorption bands are pressure broadened, and because high-opacity regions of a molecular band probe lower atmospheric pressures than do low-opacity regions, infrared observations can be used to probe the thermal structure of the atmosphere and surface of an exo-Earth. Additionally, pressure-induced absorption features can be used to indicate bulk atmospheric composition and pressure, and one such feature from N$_2$ has been detected in observations of the distant Earth near 4.3 µm (Schwieterman et al. [2015b]). These details would be true for Earth at any stage in its evolution, even in the Archean as atmospheric hazes are transparent typically transparent at infrared wavelengths. Finally, thermal infrared lightcurves could also reveal variability due to weather (and associated condensational processes), rotation, and seasons (Hearty et al. [2009], Selsis et al. [2011], Robinson [2011], Gómez-Leal et al. [2012], Cowan et al. [2012]).

Using spatially-resolved Galileo/NIMS Earth observations, along with adopted a priori knowledge of total atmospheric pressure and the CO$_2$ mixing ratio, Drossart et al. (1993) derived the thermal structures of cloud-free regions on Earth from the 4.3 µm CO$_2$ band. More recently, von Paris et al. (2013) used retrieval techniques on simulated infrared observations of a distant, modern Earth to show that low-resolution ($\lambda/\Delta\lambda = 20$) observations at signal-to-noise ratios of 10–20 could constrain thermal structure and atmospheric composition reasonably well. However, like the Drossart et al. (1993) retrievals, the results from von Paris et al. (2013) emphasize a cloud-free atmosphere. Thus, it remains unclear how realistic patchy clouds would impact our ability to understand the atmosphere and surface of an exo-Earth from thermal infrared observations.

7. SUMMARY

Exoplanetary science is rapidly progressing towards its long-term goal of discovering and characterizing Earth-like planets around our nearest stellar neighbors. We now know that exoplanets, as well as potentially habitable Earth-sized worlds, are quite common, and small worlds orbiting within the Habitable Zone of nearby cool stars have already been discovered. Advances in observational technologies, especially with regards to exoplanet direct imaging techniques,
will enable the detection of Pale Blue Dots around other stars, potentially in the not-too-distant future.

Flybys of Earth by the Galileo spacecraft in the early 1990s enabled the remote detection of habitability and life on our planet using planetary science remote sensing techniques. A combination of spatially-resolved visible and near-infrared spectral observations argued conclusively for the presence of liquid water on Earth’s surface. These same datasets indicated an atmosphere that was in a state of strong chemical disequilibrium—a sign of life—and observations at radio wavelengths contained features that indicated the presence of intelligent organisms.

More recently, a variety of observational approaches have yielded datasets that, effectively, allow us to view Earth as a distant exoplanet. While observations from spacecraft at or beyond the Moon’s orbit are ideal for understanding habitability and life signatures from the Pale Blue Dot, such data are rarely acquired. Critically, satellite and Earthshine observations complement, and fill in certain gaps between, spacecraft data for the distant Earth.

Beyond observational datasets, models have proved effective tools for simulating and characterizing Earth as an exoplanet. These tools span a wide range of complexities, including one-dimensional (vertical) spectral simulators, simple reflectance tools that capture the broadband reflectivity of Earth at visible wavelengths, and complex three-dimensional models that can simulate observations of the distant Earth at arbitrary viewing geometry across wavelengths that span the ultraviolet through the thermal infrared. Especially in the absence of frequent spacecraft observations, models of the Pale Blue Dot can serve as testing grounds for proposed approaches to detecting and characterizing Earth-like exoplanets.

Geological and bio-geochemical studies of the long-term evolution of Earth reveal a world that, while being continuously habitable and inhabited, has progressed through a strikingly wide range of surface and atmospheric states. Abundances of key atmospheric constituents, including biosignature gases, have varied by many orders of magnitude. As these gases imprint information about their concentrations on spectra of Earth, observations of the Pale Blue Dot from different geologic epochs would reveal atmospheric chemical states (and biospheres) quite distinct from modern Earth. Especially for the Archean Earth, the term “Pale Blue Dot” may not even apply.

Combining an understanding of remote sensing techniques relevant to exoplanets with knowledge of the conditions on the current and ancient Earth yields insights into approaches for detecting and studying Earth-like worlds around other stars. Broadband observations have the potential to reveal habitable environments on ocean-bearing exoplanets, and time-resolved photometry can be used to extract spatial information from spatially-unresolved data. More powerfully, spectroscopic observations at moderate resolutions can uncover key details about the surface and atmospheric state on a potentially Earth-like planet, including fundamental details relevant to life detection. Only by uncovering the key signatures that indicate the habitability and inhabitation of Earth—at any point in its evolution—can we properly design the observational tools needed to discover and fully characterize other Earths.

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