Experimental Investigation of Oxidation of Pyroxene and Basalt: Implications for Spectroscopic Analyses of the Surface of Venus and the Ages of Lava Flows

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Abstract

Alteration of basalt on the surface of Venus should be dominated by nanophase hematite and sulfate coatings, but the timescale of oxidation and how that affects spectroscopic measurements has not been extensively studied. The oxidation rate and the effect on reflectance spectroscopy are needed to constrain the age of lava flows with high emissivity that were previously measured by the European Space Agency’s Venus Express Mission. Here we investigate the effects of oxidation on the mineralogy and visible to near-infrared (VNIR) spectroscopy of augite, diopside, pyroxenite, and an alkali basalt. The experimental and spectroscopic data are used to constrain the surface basaltic weathering rates for Venus. Further, we use the spectroscopic measurements to constrain how quickly igneous signatures would degrade during oxidation of the surface. Our results show that basalt and olivine oxidize within weeks to months on the surface of Venus and their VNIR results are dominated by hematite. Pyroxenes take longer to alter with minor hematite forming on the surface—largely in cracks. Instead, Fe3+ is predominantly forming in the crystal structure. Our work suggests that if basalt on the surface of Venus contains olivine and/or glass, the high emissivity lavas previously measured would be only a few years old. If, instead, those high emissivity lavas are fully crystalline and contain no olivine or glass (which is unlikely), they would still only be decades to hundreds of years old. Therefore, this is consistent with recent work suggesting that Venus is volcanically active today.

Unified Astronomy Thesaurus concepts: Spectroscopy (1558); Venus (1763); Planetary surfaces (2113); Surface processes (2116); Volcanoes (1780)

1. Introduction and Background

The surface of Venus should be extensively altered because of its interactions with the CO2–CO–SO2 atmosphere, which is occurring at high pressures and temperatures, ~95.6 bars and ~470°C (e.g., Lodders & Fegley 1998; Seiff et al. 1985; Zolotov 2018). These gas-solid interactions should include oxidation and sulfurization, which should produce thin coatings of magnetite, hematite, and/or sulfates on the surfaces of minerals and rocks (Zolotov 2018). Previous studies have examined high-temperature oxidation of ferromagnesian silicate minerals (e.g., Fegley & Prinn 1989; Johnson & Fegley 2002; Berger et al. 2019), but more work is needed to constrain the oxidation rates of pyroxene and basalt, as well as to investigate how oxidation affects diagnostic mineral absorption bands in visible to near-infrared (VNIR) spectra.

In VNIR wavelengths, Venus’ surface can be viewed from orbit only through a few narrow spectral windows, all with wavelengths near 1 μm (1000 nm). Such visibility of Venus’ surface has been demonstrated in images from the Virtual Infrared Thermal Imaging Spectrometer instrument (VIRTIS) on the Venus Express spacecraft (Drossart et al. 2007). VIRTIS (and similar instruments) views Venus’ surface in emitted VNIR light, and properties of surface materials are retrieved by calculating its emissivity, e, by comparing the observed emission with that of a blackbody at the same temperature. Rock emissivities at Venus surface temperatures are difficult to measure (e.g., Helbert et al. 2019) but can be compared to their reflectances, r; emissivity is related to reflectance by Kirchhoff’s law, e ≃ 1 − r (Ford & Pettengill 1983). Recent work has shown that emissivities (and reflectances) near 1 μm band are related to the rock type (Gilmore et al. 2017; Dyar et al. 2019; Helbert et al. 2019; Treiman et al. 2020), primarily the iron content and oxidation state of the rock, and are largely unaffected by temperature (Treiman et al. 2020); the exception is that the spectral properties of the mineral hematite shift with increasing temperatures (Musella et al. 2002; Yamanoi et al. 2009; Treiman et al. 2020). Using these spectral windows, Smrekar et al. (2010) identified hotspots with anomalously high emissivity (low reflectance), which were interpreted to be fresh, non-weathered basaltic surfaces that may be young lava flows possibly less than 250,000 yr old. However, there was a large uncertainty in association with the ages of flows because there were no studies of how oxidation affects VNIR emissivity (and reflectance) and the timescales for these transitions.

Several earlier studies have investigated high-temperature oxidation and alteration of pyroxenes and other basaltic minerals under Venus-like conditions (e.g., Fegley & Prinn 1989; Fegley et al. 1995; Fegley & Treiman 1992; Johnson & Fegley 2002; McCanta et al. 2014; Zolotov 2018; Berger et al. 2019; Teffeteller et al. 2020) and have shown that reactions should be fairly quick and produce iron oxides and Ca-(Na) sulfates. None of these studies considered how alteration and oxidation would affect VNIR properties of the materials. Filiberto et al. (2020) investigated the oxidation of olivine at Venus surface temperatures and obtained reflectance spectra of the altered olivine. Further, Minitti et al. (2002) previously investigated the effect of oxidation on basaltic glass and
pyroxene and also obtained reflectance spectra of the altered run products, but the materials were of Martian and not
Venustian compositions; however, the results will be compared here as well. Both studies showed that Fe-bearing materials
becomes coated in hematite within days to months and that the
characteristic olivine spectral feature at 1 μm (1000 nm)
weakens and vanishes. Therefore, Filiberto et al. (2020)
suggested that if the high emissivity of lava flows (e.g.,
Smrekar et al. 2010) is caused by the presence of olivine, these
flows are, at most, a few years old.

Basalts, however, do not consist only of olivine, so it is
important to extend the work of Filiberto et al. (2020), as well
as the Martian compositions from Minitti et al. (2002), to more
realistic materials for the Venustian crust. Therefore, in this
study, we investigate the effect of oxidation on the VNIR
reflectance (and thus emissivity) of several basaltic materials:
augite, diopside, a pyroxenite, and a Venus analog basalt.
Our main goal is to constrain oxidation and weathering rates on
the surface of Venus and thus the ages of “fresh” lava flows. We
chose an alcali basalt with similar bulk chemistry to those
analyzed at Venera 13 (e.g., Kargel et al. 1993; Treiman 2007;
Filiberto 2014). We also chose three different pyroxenes
because they are abundant in terrestrial basalts, and because
Venus basalts are chemically similar to terrestrial basalts,
pyroxene should also be a major constituent of Venus basalts
(e.g., Kargel et al. 1993; Treiman 2007; Filiberto 2014).
Further, based on the mineralogy of basaltic crusts of other
planetary bodies (e.g., Tompkins & Pieters 1999; De Sanctis
et al. 2012; Ehlmann & Edwards 2014), the basaltic portion of
the crust of Venus should be dominated by pyroxenes and
similar to olivine. Pyroxenes exhibit distinctive absorption
bands near 1, as well as 2, μm (e.g., Adams 1975), the former
of which is measurable on the Venus surface by spectral
measurements from orbit (e.g., Drossart et al. 2007).

2. Starting Materials

Diopside (D), augite (A), and pyroxenite (P) samples were
purchased from Ward’s Science. The basalt (B) sample was
collected on the Sverrejell volcano (Spitsbergen, Svalbard,
Norway) in 2001 August (Treiman et al. 2002). Mineral and bulk
chemistry of each sample was verified by an electron microprobe
(EMP) analysis (Table 1). Mineral samples are relatively pure but
have minor secondary minerals. All of the samples were cut to
approximately 5 mm slabs for experimentation.

The Sverrejell basalt is a silica-undersaturated and alkalic
basalt (Table 1), which is similar to the Venera 13 analysis
(Kargel et al. 1993; Treiman 2007; Filiberto 2014). The basalt
is vesicular and contains glass, olivine, pyroxene, plagioclase,
and chromite. The porphyritic matrix contains phenocrysts of
normally zoned, euhedral-subhedral olivine with magnesian
cores (average Fo44) and abundant, normally zoned augite (a
core composition of Wo39En32Fs19). The phenocrysts are set in
glassy groundmass of plagioclase microtiles (average An49)
and euhedral-subhedral spinels, which range in composition
from Cr-magnetite to Fe-chromite.

All pyroxene samples (P, A, and D) contain large crystals. P is
an orthopyroxene dominated by orthopyroxene (OPX,
(Wo39En31Fs16) with minor clinopyroxene (CPX, Wo43En46Fs7)
and plagioclase feldspar (average An49). D is diopside
(Wo49En27Fs8). A is augite (Wo43En36Fs15) with small propor-
tions of epidote and calcite.

3. Experimental and Analytical Details

This study directly follows the methods of Filiberto et al.
(2020) for the experimental setup, reaction temperatures,
experiment durations, and atmosphere. Further, we were
guided by a similar approach of combining oxidation
experiments and VNIR analyses for Martian materials (Minitti
et al. 2002). Unmounted and unpolished samples were placed
in a Thermo Lindberg/Blue M box furnace at the NASA
Johnson Space Center (JSC), under a terrestrial atmosphere,
at 600°C, and were removed after durations of one, two, four,
and seven weeks, respectively. The experiments were conducted
in air to be directly comparable with previous work (Filiberto
et al. 2020) and to ensure enough experimental material to be
analyzable by VNIR, which has a roughly one-inch diameter
contact probe. The experiment had the oxygen fugacity of
Earth’s atmosphere, fO2 = ~0.25 bars, or ~ fayalite-
magnetite-quartz (QFM) + 10 log units (McCammon 2005).
Venus’ atmosphere has a lower fO2, at or above the magnetite-
hematite oxygen fugacity buffer (∼QFM + 5; Fegley et al.
1997). Venus’ atmosphere also differs in composition to Earth
as it primarily consists of carbon dioxide (96.5%), nitrogen
(3.5%), and minor sulfur/halogen species (Von Zahn et al.
1983). In addition, the temperature used in the experiment is

| Table 1 |
| Mineral and Bulk Chemistry of Starting Material |

| Mineral | Bulk | Olivine | CPX | Spinel (Cr) | Spinel (Ti) | CPX | Poss. Epidote | Calcite | Diopside | OPX | Plag | CPX |
|---------|------|---------|-----|------------|------------|-----|--------------|---------|----------|-----|------|-----|
| SiO2    | 43.15| 40.01   | 44.26| 0.08       | 0.13       | 53.28| 32.79       | 0.00    | 54.46    | 55.94| 53.67| 53.61|
| TiO2    | 2.02 | 0.39    | 3.91 | 5.08       | 15.25      | 0.01| 0.24       | 0.00    | 0.09     | 0.17| 0.03| 0.48 |
| Al2O3   | 15.40| 0.06    | 8.26 | 17.07      | 7.04       | 0.69| 14.29      | 0.02    | 2.44     | 1.61| 29.81| 2.19 |
| MgO     | 7.09 | 43.24   | 11.24| 6.15       | 5.50       | 12.43| 0.58      | 0.00    | 16.59    | 30.00| 0.03| 16.04|
| MnO     | 0.14 | 0.26    | 0.12 | 0.47       | 0.51       | 0.28| 0.53      | 0.02    | 0.08     | 0.21| 0.01| 0.10 |
| FeOT    | 8.64 | 16.73   | 8.38 | 45.96      | 63.72      | 9.61| 11.68     | 0.02    | 1.78     | 10.65| 0.22| 4.12 |
| Fe3O4   | 0.02 | 0.03    | 0.11 | 23.23      | 3.63       | 0.02| 0.33      | 0.01    | 0.54     | 0.23| 0.01| 0.93 |
| CaO     | 8.73 | 0.28    | 21.99| 0.18       | 0.30       | 23.39| 12.64    | 82.36   | 24.48    | 1.42 | 12.47| 22.74|
| Na2O    | 5.48 | 0.41    | 0.60 | 0.02       | 0.04       | 0.59| 0.09      | 0.01    | 0.38     | 0.02| 4.16| 0.41 |
| K2O     | 2.14 | 0.01    | 0.07 | 0.03       | 0.07       | 0.00| 0.01      | 0.00    | 0.01     | 0.00| 0.03| 0.00 |

Note. n = number of analyses. FeOT = total iron (FeO + Fe2O3). CPX = clinopyroxene. OPX = orthopyroxene. Plag = plagioclase.
slightly higher than the present-day surface temperature of Venus, ∼470°C (Seiff et al. 1985). In Filiberto et al. (2020), we showed that this experimental approach produced similar results (both the mineralogy and rate of oxidation) to experiments run under Venus-like atmospheric compositions (Berger et al. 2019; Teffeteller et al. 2020). Further, as shown below, our experimental results produce mineralogical changes on the same rate of oxidation as previous studies (Berger et al. 2019; Filiberto et al. 2020; Teffeteller et al. 2020). Therefore, although these differences are indeed potential limitations in the presented work, the experimental results are applicable for studying alteration on the surface of Venus.

Mineral chemical analyses were obtained by EMP analysis using the Cameca SX 100 instrument at the JSC from polished thin sections of unoxidized samples. We used an electron accelerating potential of 15 kV with a defocused beam for the basaltic sample and a focused beam for all pyroxene mineral analyses. Beam currents were set at 1 and 20 nA, depending on the sample. The instrument was calibrated for Na, Mg, Al, Si, K, Ca, Ti, Fe, Cr, and Mn, using natural and synthetic mineral standards. Analyses with sums totaling <99.5% and >100.99% for pyroxene, olivine, and plagioclase were discarded. For alteration phases and the bulk chemical analyses of the basalt, all analyses were included.

VNR reflectance spectra of cut, flat surfaces of unoxidized and oxidized samples were obtained using a Spectral Evolution OreXpress Spectrometer with a Benchtop Reflectance Probe. Reflectances were measured between 0.35 and 2.5 μm (350 and 2500 nm) wavelengths. For the oxidized samples, the reacted surfaces (i.e., that faced upward in the box furnace) were analyzed. A standard white reflectance panel, supplied by Spectral Evolution, was used to correct the measurements taken.

The reacted surface of each sample was also examined by Scanning Electron Microscopy (SEM) using the JEOL JSM-5910LV at the JSC. Unmounted samples were carbon coated and taped to aluminum disks. Secondary electron images (SEI) and back-scattered electron (BSE) images were obtained at a 15 kV accelerating potential and working distances between 18 and 23 mm.

4. Results

4.1. Alkali Basalt

The reflectance spectrum of the unoxidized SverreFjell basalt is typical of a basalt (Figure 1(a))—low overall reflectance (∼5%) with a blue continuum slope (decreasing reflectance with an increasing wavelength). Two small absorption features can be discerned at 1.0 and 2.2 μm; these bands are consistent with high calcium pyroxene (Cloutis & Gaffey 1991) and likely arise from augite, which is the most abundant mineral in the basalt. Our experimental alteration of the basalt caused profound changes in its reflectance spectrum (Figure 1(A)), which are broadly consistent with experimental results for oxidation of basaltic glass of Martian composition (Miniti et al. 2002). After only one week of oxidation, an absorption edge appeared between ∼430 and 460 nm, accompanied by an increase in reflectance at wavelengths longer than 460 nm. With increasing duration of oxidation, the maximum reflectance (∼450 nm) shifts to longer wavelengths (eventually to ∼830 nm) and the overall reflectance increases (from 5% to 15%; Figure 1(A)).

These changes are consistent with the formation and thickening of a coating of hematite on the basalt surface, with the final (week seven) spectrum being essentially that of pure hematite (Morris et al. 1985; Yamanoi et al. 2009). Furthermore, the hematite appears more consistent with nanophase hematite than crystalline (specular) hematite due to the absence of the 860 nm absorption band (Morris et al. 1989). Aside from the reflectance peak, the 1 μm band, related to pyroxene, disappears after two weeks of oxidation, as does a weak band at 680 nm (possibly from a Fe³⁺ crystal-field transition; Sherman et al. 1982).

The change in the basalt’s reflectance spectrum is echoed by visual and SEM examination of its surfaces. The weathered basalt surfaces become visually redder (not to be confused with a reddened continuum) with the duration of oxidation (Figure 1(A)), consistent with an ever-increasing surface coverage by fine-grained (pigmentary) hematite (e.g., Morris et al. 1989; Lucey et al. 2018). In BSE images of the surface, fine-grained (nanophase) hematite appears at the shortest weathering duration and becomes more abundant as weathering duration increases. The abundance of nanophase hematite is greatest on the olivine phenocrysts, the glass, and the ferroan rims on the augite phenocrysts. Interestingly, VNR spectra are highly sensitive to small changes in the amount of hematite present on sample surfaces (Figure 2), as hematite is forming on the surface but never fully coats the entire chip. Filiberto et al. (2020) noted something similar with the olivine oxidation experiments where Raman analyses were consistent with samples dominated by olivine, while the VNR results were dominated by hematite even before a full coating developed. Similarly, our most weathered sample, seven weeks, is not completely coated with hematite, even though its reflectance spectrum is consistent with near-pure hematite (Morris et al. 1985).

4.2. Diopside

The reflectance spectrum of the unoxidized diopside sample (D) is typical for Ca-clino.pyroxenes (Figure 1(B); Schade et al. 2004; Klima et al. 2011a, 2011b). The two prominent reflectance minima, near ∼1100 and ∼2300 nm, are caused by crystal-field transitions of Fe²⁺ in the two crystallographic sites (Hazen et al. 1978; Burns 1993). The weak absorption band at ∼800 nm is attributable to Fe²⁺ → Fe³⁺ intervalence charge transfer (IVCT; Straub et al. 1991; Burns 1993).

After oxidation, the overall spectral signal of the diopside was only slightly modified. The overall reflectance increases slightly with increased oxidation, which probably represents the increasing proportion of (relatively brighter) hematite. Another observation is that the band depths of all spectral features increase with increasing oxidation. After seven weeks of oxidation, the 1.1 μm (1100 nm) band depth (BD) increases by 57% with an initial BD = 0.34 compared with to the seven week oxidized sample at BD = 0.70. The 2.3 μm (2300 nm) BD intensity increases by 45% with an initial BD = 0.05 in contrast to the seven week oxidized sample with a BD = 0.09. In addition, we find that both characteristic pyroxene bands, at ∼1.1 and ∼2.2 μm, are shifted to slightly longer wavelengths. To shift these two bands to longer wavelengths, the crystallographic sites in the pyroxene need to contain a larger cation, such as iron or calcium (e.g., Burns 1993). Because the proportion of Ca is not increasing in the pyroxene and the iron is lost to the hematite layer, we suggest that the iron in the
Pyroxene is migrating to the surface, resulting in an appearance of an increasing abundance of iron in the pyroxene. The increasing 1 and 2 μm absorption strengths further support this interpretation.

An alternative explanation is that the cations in the starting diopside are significantly disordered among their sites (e.g., having cooled rapidly from a high temperature). The high temperature of our experiments enabled the cations to move to a more ordered configuration, i.e., concentrating Fe^{2+} from the M1 to the M2 sites (e.g., Burns et al. 1991; Trang et al. 2013). As a result of increasing the order within the pyroxene crystallographic structure, the Fe^{2+} migrates to the M2 sites, allowing for the 1 and 2 μm absorption features to shift to longer wavelengths and to have larger band depths. Future experiments will be necessary to determine which process is at work.

During oxidation, the principal changes in diopside’s reflectance spectrum are in the <1000 nm region, which can be assigned to a newly formed nanophase hematite (the shoulder is near 700 nm, the absorption band is near 860 nm; Morris et al. 1985) form. The reflectance associated with the formation of hematite slightly increases as it shifts to longer wavelengths with the increasing time of oxidation. Further, the Fe^{2+} → Fe^{3+} IVCT near 790 nm becomes more pronounced with longer durations of oxidation. Straub et al. (1991) suggested that this change represents oxidation of Fe^{2+} to Fe^{3+} in the pyroxene structure rather than formation of nanophase hematite. Small proportions of hematite did form, as shown in Figure 1.

Figure 1. Reflectance spectra of samples before and after oxidation at 600°C: (A) alkali basalt, (B) diopside, (C) augite, (D) orthopyroxenite, and (E) olivine, modified from Filiberto et al. (2020).
by the BSE images of the reacted surfaces (Figure 3). The small changes in mineralogy and spectra are consistent with the low-iron bulk chemistry of the diopside, which only has 1.78 wt. % FeO.<sub>T</sub>.

### 4.3. Orthopyroxenite

The reflectance spectrum of the unoxidized orthopyroxenite sample, P, (Figure 1(D)) shows the typical two broad absorption bands for pyroxene, centered at ~0.95 and ~1.9 μm, similar to natural pyroxenes (Adams 1975); these are attributed to Fe<sup>2+</sup>-crystal-field transitions in crystallographic sites (Besançon 1981). The oxidized samples show a gradual narrowing in the 950 nm band and an approximate 20% increase in band depth (initial BD = 0.4 compared with week seven’s BD = 1.88) by the end of the seven weeks (Figure 1(D)). The overall reflectance increases slightly with increased oxidation, which probably represents the increasing proportion of (relatively brighter) hematite. As with the diopside, the most significant changes are in the shorter wavelengths (i.e., visible) where the absorption peak becomes more prominent, consistent with the formation of hematite. The BSE images also show a gradual increase in the

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**Figure 2.** BSE images of the unoxidized (A) compared with oxidized alkali basalt samples for: (B) one week, (C) two weeks, (D) four weeks, and (E) seven weeks of oxidation. The unaltered basalt is dominated by pyroxene (gray) with euhedral to skeletal olivine phenocrysts. The sample also contains a pocket of glass that is partially devitrified and vesicles. Bright nearly square shaped crystals are spinel. Finally, plagioclase is present in small amounts (darkest spots). In the altered samples, nanophasic hematite are shown by the small white specks, which increase with the time of oxidation. The small bright spots are concentrated on material that was previously interstitial glass, olivine phenocrysts, and rims of pyroxene.
amount of hematite on the surface, with the hematite forming preferentially near cracks (Figure 4).

4.4. Augite

The reflectance spectrum of the augite sample (A) is quite different from those of the D and P pyroxenes: it has a very broad reflectance minimum between 0.5 and $\sim$2.5 $\mu$m (Figure 1(C)) but over a range of small reflectances $<$5%. This reflectance spectrum is similar to the “type A” augites of Adams (1975) and Schade et al. (2004). This broad absorption is due to the augite having high abundances of Ca$^{2+}$, which occupies all or nearly all of the M2 crystallographic sites. Because Fe$^{2+}$ cannot be in the M2 sites, the 1 and 2 $\mu$m absorption bands are either absent or severely weakened. Consequently, Fe$^{2+}$ occupies the smaller M1 crystallographic sites, which produces two weak absorptions at 1 $\mu$m, which is not very apparent in the spectra of this sample due the low Fe$^{2+}$ abundance within the sample (Table 1). Regardless of the placement of the Fe$^{2+}$ in the pyroxene crystal structure, oxidation of the augite causes changes similar to those seen in the orthopyroxene and diopside samples: overall reflectance
increases, and typical hematite signatures become apparent only at the shortest wavelengths (Figure 5).

5. Discussion

The experimental results here, combined with previous studies (Berger et al. 2019; Filiberto et al. 2020) and the similar results for Martian materials (Minitti et al. 2002), can be used to constrain oxidation rates of basalt and basaltic minerals on Venus, which in turn can be used to constrain the ages of lava flows on Venus. Thus, we can calibrate the ages of lava flows that have been measured with near-infrared (NIR) emissivity, e.g., as done by Smrekar et al. (2010) and D’Incecco et al. (2017) using data from VIRTIS (Drossart et al. 2007).

The basalt sample became coated with hematite, to the point of obscuring its VNIR reflectance, in approximately the same time as olivine and glass (Figure 1), after a month of reaction (Minitti et al. 2002; Filiberto et al. 2020). Further, the basalt sample formed hematite coatings on a similar timescale as basaltic glass and basaltic minerals under a CO$_2$–CO gas (e.g., Berger et al. 2019; Teffeteller et al. 2019, 2020). Therefore, this implies that the rate of hematite formation is essentially independent of oxygen fugacity or the difference in atmospheric conditions, which implies that the rate-limiting
mechanism in oxidation is not the availability of oxygen at the surface but the diffusion of iron to the rock surface as was previously suggested (Cooper et al. 1996; Filiberto et al. 2020). Together, these works suggest that alteration on Venus is dominated by the rapid development of thin coatings ($<1 \mu m$) of nanophase hematite at its surface. This is consistent with the red color of Venus’ surface rock and regolith at the Venera 9 and 10 landing sites, which is suggestive of (pigmentary or nanophase) hematite (Pieters et al. 1986). However, sulfur may change alteration rates and will likely cause sulfates, in addition to hematite, to form on the surface of basalt (e.g., Fegley et al. 1995; Fegley & Treiman 1992; Johnson & Fegley 2002; Zolotov 2018; Berger et al. 2019).

In our experiments, pyroxenes are altered and oxidized more slowly than olivine or basalt consistent with previous experimental results for Mars (Minitti et al. 2002). The VNIR reflectances of pyroxenes increase moderately across all wavelengths longer than $\sim 800$ nm, presumably because the hematite coatings have higher reflectance than the underlying pyroxenes (Figure 1), as previously suggested (Minitti et al. 2002). Reflectances of the augite and orthopyroxenite samples changed more than the diopside, which had the lowest iron

Figure 5. BSE images of the unoxidized (A) compared with oxidized augite samples for: (A) one week, (B) two weeks, (C) four weeks, and (D) seven weeks of oxidation. The unoxidized augite sample, similar to the previous ones, also shows the saw cuts through it; however, unlike the previous ones, it also shows dark spots, which represent the alteration materials (see Table 1). With progressive oxidation, bright white specks show up on the sample, which represent the nanophase hematite formation. As with the other pyroxene samples, the hematite is concentrated on cracks, inclusions, and saw cuts. Panel (B) shows hematite formation on inclusions and (D) shows hematite formation on the saw cuts.
content. Unsurprising, this suggests that iron-bearing pyroxene will oxidize faster than low-iron pyroxene. Spectrally, all of the pyroxenes would be detectable by VNIR reflectance (and emissivity) on the surface of Venus, even after two months of alteration. Comparing this with the recent work on olivine oxidation (Filberto et al. 2020) shows that pyroxene alteration is about an order of magnitude slower than olivine oxidation. Projecting this out, assuming this rate does not significantly change, would suggest that iron-bearing pyroxenes would be covered with hematite within decade timescales, with the caveat that sulfur in the atmosphere would likely increase the alteration rate (e.g., Fegley & Treiman 1992; Berger et al. 2019).

Using these new alteration rates and the rate at which alteration mineral formation affects VNIR spectroscopic signatures, we can now revise the timescales of volcanism on Venus. Multiple areas have been identified as regions with active volcanism (e.g., Smrekar et al. 2010; Shalygin et al. 2015; D’Incecco et al. 2017). The volcanic rises of the Imdr, Themis, and Dione Regiones with high emissivity (low reflectance) have been correlated to signifying fresh basaltic surfaces, and the lava flows were calculated with an age range of 250,000 ka to 2.5 Ma (Smrekar et al. 2010). Filiberto et al. (2020), based on rates of olivine oxidation, suggested that these flows were, at most, a few years old. Combining our new experimental results, with the previous results on olivine only, shows that basalt oxidizes at rates similar to those of olivine alone. Based on the experimental time it took for a basalt (containing glass, olivine, and pyroxene) to become oxidized to the point where pyroxene absorption bands are absent, we determined that these lava flows on Venus are likely only a few years old; however, if the basalt is fully crystalline and dominated by pyroxene, with no glass or olivine, these lava flows would be about an order of magnitude older on the timescale of tens to hundreds of years old. Both estimates suggest that these lava flows are in fact not only geologically young, as was previously suggested by D’Incecco et al. (2017) and Smrekar et al. (2010), but are in fact present-day volcanism and these flows may have contributed to the atmospheric SO$_2$ plumes detected by the Pioneer Venus and Venus Express orbiters (Esposito 1984; Marcq et al. 2013). Further, it has been suggested by radar investigations that other lava flows, such as as at Maat Mons, may have even higher emissivities/reflectances and, as such, may be similar or even younger in age, but there are no VNIR emissivity data (i.e., from Venus Express) for these flows (Brossier et al. 2020). Previous work has suggested, based on Monte Carlo simulations, that Venus was volcanically resurfaced around 300 m.y. Thereafter, lava production rates fell to an estimated rate over the last 10 m.y. of 0.01–0.15 km$^2$ yr$^{-1}$, which is less than Earth’s current rate of intraplate volcanism (e.g., Strom et al. 1994). If the implications of our alteration experiments are correct, volcanic rises within Imdr, Themis, and Dione Regiones, as well as possibly Maat Mons and others not previously analyzed, are producing magma today at rates similar to terrestrial hot spot volcanism. This would suggest that each volcanic center is producing approximately up to 0.1 km$^2$ yr$^{-1}$ based on terrestrial averages for basaltic hot spot volcanoes from Crisp (1984); however, without a full estimate of the number of active volcanoes on the Venus surface from global high-resolution VNIR measurements, it is impossible to fully constrain the global lava production rate using this method.

Alteration experiments combined with VNIR as described here can provide constraints for detecting areas on Venus characterized by such recent eruptions. Therefore, for future missions, this implies that orbital emissivity measurements in the 1 μm region can be used to calculate not only the iron content and oxidation state of the surface of Venus (e.g., Dyar et al. 2019), but these results can be used to estimate the ages of basalt lava flows. Future proposed missions, such as EnVision; Deep Atmosphere Venus Investigation of Noble gases, Chemistry, and Imaging + (DAVINCI+); or Venus Emissivity, Radio Science, InSAR, Topography, and Spectroscopy (VERITAS; Ghail et al. 2012, 2018; Glaze et al. 2017, 2018; Garvin et al. 2020; Smrekar et al. 2020; Widemann et al. 2020), should be able to see changes in these flows consistent with oxidation and alteration.

6. Summary

Our experimental results confirm lander evidence and theoretical inferences that basaltic rock on Venus’ surface should be coated by hematite (e.g., Pieters et al. 1986; Fegley et al. 1995). Further, the rate of oxidation and alteration is rapid; olivine and basalt oxidizes within weeks to months, causing the VNIR optical characteristics of basalt and minerals (olivine, pyroxene) to not be detectable after a few decades. Our results lend support to the recent inference (Filiberto et al. 2020) that lava flows at Idunn Mons (Smrekar et al. 2010; D’Incecco et al. 2017) are in fact only years to decades old. Ongoing and future experiments under atmospheres more like Venus’ own (e.g., at lower $f$O$_2$ and containing sulfur-bearing gas species) should further constrain these timescales. Finally, future Venus spacecraft should incorporate imaging spectrometers, sensitive at the wavelengths of the “windows” through Venus’ atmosphere (in the 1 μm region), to constrain the iron contents, oxidation states, and thus ages of Venus’ lava flows.

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