

Interactive comment on “Synoptic Ozone, Cloud Reflectivity, and Erythemal Irradiance from Sunrise to Sunset for the Whole Earth as viewed by the DSCOVR spacecraft from Lagrange-1” by Jay Herman et al.

Jay Herman et al.

jay.r.herman@nasa.gov

Received and published: 28 October 2017

I have changed the title to

EPIC Synoptic Ozone, Cloud Reflectivity, and Erythemal Irradiance from Sunrise to Sunset for the Whole Earth as viewed by the DSCOVR spacecraft from the Earth Sun Lagrange-1 Orbit

Interactive comment on Atmos. Meas. Tech. Discuss., doi:10.5194/amt-2017-155, 2017.

[Printer-friendly version](#)

[Discussion paper](#)



Interactive comment on “Synoptic Ozone, Cloud Reflectivity, and Erythemal Irradiance from Sunrise to Sunset for the Whole Earth as viewed by the DSCOVR spacecraft from Lagrange-1” by Jay Herman et al.

Jay Herman et al.

jay.r.herman@nasa.gov

Received and published: 19 October 2017

Response to Referee #2

I have modified the abstract to reflect the reviewer's comments

EPIC (Earth Polychromatic Imaging Camera) onboard the DSCOVR (Deep Space Climate Observatory) spacecraft is the first Earth science instrument located near the Earth-Sun gravitational plus centrifugal force balance point, Lagrange-1. EPIC measures Earth reflected radiances in 10 wavelength channels ranging from 317.5 nm to

[Printer-friendly version](#)

[Discussion paper](#)



779.5 nm. Of these channels, four are in the UV range 317.5, 325, 340, and 388 nm, which are used to retrieve O₃, 388 nm scene reflectivity (LER Lambert Equivalent Reflectivity), SO₂, and aerosol properties. Unlike low earth orbiting satellite instruments near noon values, these synoptic quantities for the entire sunlit globe from sunrise to sunset obtained every 68 minutes when it is summer or 110 minutes in winter at the receiving antenna in Wallops Island, Virginia. Depending on solar zenith angle, either 317.5 or 325 nm channels are combined with 340 and 388 nm to derive ozone amounts. As part of the ozone algorithm, the 388 nm channel is used to derive LER. The retrieved ozone amounts and LER are combined to derive the Erythema irradiance for the entire sunlit Earth's surface, 2048x2048 points, at a nadir resolution of 18 x 18 km² using a computationally efficient approximation to a radiative transfer calculation of irradiance. Corrections are made for altitude above sea level and for the reduced transmission by clouds based on retrieved LER.

Also highlighting the value of EPIC measurements for the general public (UV index – is this passed on? Published anywhere else?) would be an interesting addition. I have given the algorithm to responsible parties at NOAA. They expressed interest, but have not indicated that they are using the new approach.

Please include explanations of abbreviations in the abstract. Done

p.4, line 11 ff: This paragraph fits better further up in the introduction.

I moved the paragraph into the introduction. The data and images of the changing synoptic cloud cover from sunrise to sunset are unique to the EPIC satellite instrument. Neither geostationary nor low earth orbiting satellites can produce these data or images. Geostationary satellites could produce something similar, but to date, none have the UV channels for ozone and LER, and geostationary satellites are limited to a range of approximately $\pm 60^\circ$ latitude and $\pm 60^\circ$ longitude. While low earth orbiting satellite data can be combined to produce a global representation of ozone and cloud cover, all the ozone and cloud cover are for a fixed local time (e.g., 13:30 hours for OMI) and are

[Printer-friendly version](#)[Discussion paper](#)

not representative of the atmosphere at other times of the day. 1.1 EPIC Instrument
p.5, line 187: How can errors cancel each other out? Please explain.

These are not errors. Rather to form the albedo, one takes the ratio of two quantities that both contain the solar Fraunhofer line structure. The resulting ratio does not have the Fraunhofer line structure. I modified the related sentence to read:

Because the albedo spectra AM (Eq. 1) removes the Fraunhofer line structure contained in both the solar irradiance SM and the reflected Earth radiance IM, the interpolation and convolution of AM has better accuracy than directly using IM.

p.10, line 335: Any ideas where these difference come from?

The differences arise from errors in the various satellite's calibrations.

F01: air and vacuum WL in same Figure is confusing. I chose to provide the original laboratory data (curves) done in air. I shifted the central wavelength to vacuum, since those are the values that are used in all of the science. I have added a note in the caption to clarify.

f01 Filter transmission functions (percent) for the 10 EPIC wavelengths based on laboratory measurements done in air. The central wavelength label is the shifted value used for the instrument in the vacuum of space.

F11: change red trace to grey/black or explain My error. I have fixed the figure so that all are black

p.2, line 33: orbit around

I am not sure of the reviewer's reference. The text says , "to an orbit near the Earth-Sun gravitational plus centrifugal force balance point". The orbit is a Lissajous figure about the L-1 point. At times the shape of the orbit is an ellipse with the L-1 point at a focalpoint, other times the orbit is a circle with the L-1 point at the center. The orbit goes from elliptical to a circle and back to elliptical every 5 years. However, this

[Printer-friendly version](#)[Discussion paper](#)

periodicity is altered unpredictably by in-orbit thruster corrections to account for lunar perturbations and to prevent the spacecraft from leaving the quasi-stable L-1 orbit.

The text now reads, “The DSCOVR (Deep Space Climate Observatory) spacecraft was successfully launched on 11 February 2015 to a lissajous figure orbit near the Earth-Sun gravitational plus centrifugal force balance point, Lagrange-1 (L-1), 1.5x10⁶ km from the Earth.”

p.2, line 36: optimized

The orbit was selected for earth observations, and so just turned out to be optimum for early solar storm warnings. For clarity, I changed the sentence to read, “The DSCOVR mission at L-1 is at an optimum location for early warning”

p.2, line 62: or over ice

Fresh snow over ice is correct

It turns out that the most reflective scenes are cloud-free skies with fresh snow deposited over ice. Fresh snow over land is usually not as bright because of photons lost to the absorbing ground, and scenes with only ice have a lower reflectivity because ice surfaces are usually older and darkened by pollution.

p.3, line 85: 10 wavelength Each of the 10 wavelength measurements

p.3, line 85: at slightly I do not understand the reviewer’s objection. The sentence reads, “Each of the 10 wavelength measurements is obtained at slightly different times.”.

p.4, line 120: and are not OK (see above on page 1)

p.5, line 158: result in large OK

Section 6: several inconsistencies in use of capital “E” or not in erythemal The word “erythemal” is now all lower case except at beginnings of sentences or when combined

[Printer-friendly version](#)[Discussion paper](#)

1 Synoptic Ozone, Cloud Reflectivity, and Erythemal Irradiance from Sunrise to Sunset for the Whole Earth
2 as viewed by the DSCOVR spacecraft from the Earth Sun Lagrange-1 Orbit

3 Jay Herman¹, Liang Huang², Richard McPeters³, Jerry Ziemke³, Alexander Cede⁴, Karin Blank³

4 Abstract

5 EPIC (Earth Polychromatic Imaging Camera) onboard the DSCOVR (Deep Space Climate Observatory)
6 spacecraft is the first Earth science instrument located near the Earth-Sun gravitational plus centrifugal
7 force balance point, Lagrange-1. EPIC measures Earth reflected radiances in 10 wavelength channels
8 ranging from 317.5 nm to 779.5 nm. Of these channels, four are in the UV range 317.5, 325, 340, and
9 388 nm, which are used to retrieve O₃, 388 nm scene reflectivity (LER Lambert Equivalent Reflectivity),
10 SO₂, and aerosol properties. Unlike low earth orbiting satellite instruments, these quantities are derived
11 synoptically at Goddard Space Flight Center for the entire sunlit globe from sunrise to sunset every 68
12 minutes when it is summer or 110 minutes in winter at the receiving antenna in Wallops Island, Virginia,
13 respectively. Depending on solar zenith angle, either 317.5 or 325 nm channels are combined with 340
14 and 388 nm to derive ozone amounts. As part of the ozone algorithm, the 388 nm channel is used to
15 derive LER. The retrieved ozone amounts and LER are combined to derive the Erythemal irradiance for
16 the entire sunlit Earth's surface, 2048x2048 points, at a nadir resolution of 18 x 18 km² using a
17 computationally efficient approximation to a radiative transfer calculation of irradiance. Corrections are
18 made for altitude above sea level and for the reduced transmission by clouds based on retrieved LER.

19

20

21

22 ¹University of Maryland Baltimore County, Maryland

23 ²Science Systems and Applications, Lanham, Maryland

24 ³NASA Goddard Space Flight Center, Greenbelt, Maryland

25 ⁴SciGlob Instruments and Services, Maryland

26

27

28 DSCOVR/EPIC Synoptic Ozone, Cloud Reflectivity, and Erythral Irradiance From Sunrise to Sunset for
29 the Whole Earth as viewed from an Earth-Sun Lagrange-1 Orbit

30 **1.0 Introduction**

31 The DSCOVR (Deep Space Climate Observatory) spacecraft was successfully launched on 11
32 February 2015 to an orbit near the Earth-Sun gravitational plus centrifugal force balance point,
33 Lagrange-1 (L-1), 1.5×10^6 km from the Earth. The earth pointing instruments on the DSCOVR spacecraft
34 placed in orbit about the L-1 point will simultaneously observe the sun illuminated earth's disk from
35 sunrise to sunset. An illustration of the orbit is given in the Appendix (see <https://epic.gsfc.nasa.gov> for
36 details). DSCOVR started to transmit Earth data after it achieved a quasi-stable orbit in mid-June 2015.
37 The DSCOVR mission at L-1 is optimum for early warning solar flare observations (magnetic field,
38 electron, and proton fluxes) from instruments contained on the sunward side of DSCOVR, and contains
39 two Earth-viewing instruments allowing continuous observation of the sunlit face of the Earth. The EPIC
40 (Earth Polychromatic Imaging Camera) instrument onboard DSCOVR images the Earth in ten narrow
41 band wavelength channels (up to 2048 x 2048 pixels), producing both color images of the Earth and
42 science data products such as ozone, SO₂, aerosol amounts, cloud reflectivity, UV surface irradiance,
43 cloud and aerosol heights, and vegetation indices. This paper discusses the UV science products O₃,
44 cloud reflectivity, and UV surface irradiance, methods of retrieval, and EPIC's UV in-flight calibration.

45 **The data and images of the changing synoptic cloud cover from sunrise to sunset are unique to**
46 **the EPIC satellite instrument. Neither geostationary nor low earth orbiting satellites can produce these**
47 **data or images. Geostationary satellites could produce something similar, but to date, none have the UV**
48 **channels for ozone and LER, and geostationary satellites are limited to a range of approximately $\pm 60^\circ$**
49 **latitude and $\pm 60^\circ$ longitude. While low earth orbiting satellite data can be combined to produce a global**
50 **representation of ozone and cloud cover, all the ozone and cloud cover are for a fixed local time (e.g.,**
51 **13:30 hours for OMI) and is not representative of the atmosphere at other times of the day.**

52 **1.1 EPIC Instrument**

53 The EPIC instrument consists of a 30-cm aperture 283.642 cm focal length Cassegrain telescope
54 containing a multi-element field-lens group focusing light onto a UV sensitive 2048 x 2048 hafnium
55 coated CCD detector with 12 bit readout electronics. Images are made through ten narrow-band filters,
56 four in the ultraviolet, four in the visible, and two in the near infrared. The 10 filter transmission
57 functions are shown in Fig. 1. Observations are made as light passes sequentially through each of ten
58 narrow-band filters mounted in two moveable filter wheels and through an exposure control 3-slot
59 rotating shutter. The exposure times for each wavelength were adjusted in-flight to achieve an
60 approximately 80 % CCD electron well fill in the brightest scenes, which were observed during the first
61 week of operation, to avoid saturation and leaking from one pixel to another (blooming). Earth exposure
62 times range from about 654 milliseconds at 317.5 nm to 22 milliseconds at 551 nm, which have not
63 changed during the current life of the mission. Another set of exposure times was determined for
64 viewing the full moon as seen from the Earth (Table 1). The CCD has a well depth of approximately
65 8.5×10^4 electrons (a maximum signal to noise ratio SNR of 290:1) before a small dark current correction
66 that is a function of its in-flight operating temperature of -20°C. The 12-bit readout means that there

67 are 2^{11} (2048) readout steps or counts (42 electrons/count). The counts divided by the exposure time
68 (counts/second) are converted to radiances or albedos using in-flight scene matching calibration from
69 low earth orbit satellites (see Sect. 1.2 and Table 2). The maximum SNR applies to the brightest of
70 scenes over high clouds or fresh snow over ice. Cloud-free and snow-free scenes have much lower SNR,
71 which affects the visible channels more than the UV channels because of the lower scene contrasts with
72 clouds caused by enhanced UV Rayleigh scattering. There are occasional bright flashes caused by ice
73 crystals in high clouds that saturate a few pixels (see Fig. 2 and Marshak et al., 2017) in the equatorial
74 and mid-latitude regions.

75 The filters of interest for calculating ozone amounts, aerosol index, and cloud reflectivity are
76 centered on 317.5, 325, 340, and 388 nm in the wavelength band with full widths at half maximum
77 (FWHMs) 1.0, 1.0, 2.7, and 2.6 nm, respectively. For the UV channels, 2 x 2 individual pixels are
78 averaged onboard the spacecraft to yield an effective 1024 x 1024 pixel image corresponding to an 18 x
79 18 km² resolution at the observed center of the Earth's sunlit disk. The effective spatial resolution
80 decreases as the secant of the angle between EPIC's sub-earth point and the normal to the earth's
81 surface. Only the 443 nm channel is retrieved at full resolution to help with resolving cloud cover and
82 obtaining improved color images. The sampling resolution of a single pixel is about 8 x 8 km² (about 1
83 arcsecond), but including the effect of the optical point-spread function, the effective 443 nm channel
84 resolution is about 10 km. The effective resolution at 443 nm has been verified by looking at clear
85 scenes over the Nile River in Egypt and, occasionally, the cloud-free Amazon River in Brazil.

86
87 EPIC data has been obtained since June 15, 2015 at a rate of one set of 10 wavelengths every 68
88 minutes during Northern Hemisphere (NH) summer and one set every 110 minutes in the winter. The
89 difference between summer and winter rates is caused by the reduced number of hours in the winter
90 when the antenna (located at Wallops Island, Virginia) is in view of the spacecraft, and limitations from
91 the spacecraft memory technology from the late 1990s.

92
93 Each of the 10-wavelength measurements is obtained at slightly different times. The first filter in
94 the sequence is 443 nm, which takes about 2 minutes to complete a measurement (28 ms exposure
95 time (Table 1) plus CCD readout and onboard processing time that includes 12-bit jpeg compression of a
96 2048 x 2048 pixel image). The remaining 9 filter measurements take a total of about 5 minutes
97 (exposure times plus CCD readout into memory) and then another 13 minutes to process the data for
98 the 9 filters (this includes 12-bit jpeg compression of 1024 x 1024 images that have been averaged
99 onboard in groups of 2x2 pixels before compression). Adjacent pairs of wavelengths are measured at 30
100 second intervals before the onboard processing is started. This means the individual channel images are
101 not co-located at the pixel level because of earth rotation (15.03° per hour or about 1670 km per hour
102 at the equator), the slow rotation of the spacecraft, 0.082° per hour, and a small amount of spacecraft
103 jitter). Each pixel views about 1 arc second or 2.78×10^{-4} degrees. Data from an onboard star-tracker and
104 feedback from the earth's image on the CCD keep the images approximately centered on the CCD. The
105 lack of native channel-to-channel colocation requires an elaborate spherical geometry geolocation
106 analysis to adjust the data to a common latitude x longitude grid with an accuracy of 1/4 of a pixel.

107 A description of the EPIC instrument, its orbit, and some of the data products can be obtained
 108 from http://avdc.gsfc.nasa.gov/pub/DSCOVER/Web_EPIC/ and from <http://epic.gsfc.nasa.gov/>. The EPIC
 109 raw counts/second and science data (Version 2 used in this paper) are archived at
 110 https://eosweb.larc.nasa.gov/project/dscovr/dscovr_table in HDF5 format.

111
 112 This paper presents examples of the ozone and scene reflectivity retrievals that are used to
 113 obtain unique estimates of erythemal UV irradiance (or UV Index, UVI) as a function of latitude,
 114 longitude, local solar time (LST), and altitude above sea level (ASL). Since this is the first paper on EPIC
 115 retrieved ozone, Sect. 1 contains a brief description of the calibration of the four UV channels and the
 116 ozone retrieval algorithm. Sect. 2 shows examples of natural color images, Sect. 3 gives an example of
 117 retrieved ozone and the corresponding 388 nm Lambert Equivalent Reflectivity (LER, Herman et al.,
 118 2009), Sect. 4 presents a validation of EPIC retrieved ozone compared to ozone from ground-based and
 119 satellite data, Sect. 5 shows details of the latitudinal and longitudinal synoptic variability of ozone, and
 120 Sect. 6 presents new results showing the sunrise to sunset variability of UV erythemal radiation reaching
 121 the Earth’s surface including the reduction by clouds from sunrise to sunset.

122 1.2 Calibration

123 Before the raw EPIC data (counts per second) can be used, a number of pre-processing steps
 124 must be accomplished. The major steps are 1) measuring and subtracting the dark current signal, 2)
 125 “flat-fielding” the CCD so that the sensitivity differences between all four million pixels are determined
 126 and corrected, 3) correcting for stray-light effects to account for light that should be going to a particular
 127 pixel, but instead is scattered to different pixels, and 4) determining the radiometric calibration for each
 128 wavelength channel in terms of EPIC counts/second to be converted to earth normalized radiances or
 129 reflectances (backscattered at approximately 172°). The earth upwelling normalized radiance I_M ($W/(m^2$
 130 $nm\ sr)$) at the top of the atmosphere (TOA) is defined in terms of the albedo A_M given by Eq. 1,

$$A_M = \frac{I_M}{S_M/D_E^2} \quad (sr^{-1}) \quad (1)$$

131 for wavelength bands $M=1$ to 4, S_M is the incident solar irradiance ($W/(m^2\ nm)$) weighted with the filter
 132 function for band M at 1 AU and D_E is the sun-earth distance in AU (astronomical units). Since EPIC does
 133 not measure solar irradiance, we use a high resolution solar irradiance spectrum, $S(\lambda)$ (Dobber et al.,
 134 2008), as a reference solar spectrum. The reference spectrum is weighted with EPIC’s filter transmission
 135 functions $T_M(\lambda)$ (Fig. 1) to obtain each EPIC channel’s weighted solar irradiance S_M at solar-earth
 136 distance at 1 astronomical unit (Eqs. 1 and 2).

$$S_M = \int_{\lambda_1}^{\lambda_2} T_M(\lambda)S(\lambda)d\lambda / \int_{\lambda_1}^{\lambda_2} T_M(\lambda)d\lambda \quad (Wm^{-2}nm^{-1}) \quad (2)$$

137

138 In-flight radiometric calibration is accomplished by comparison with albedo values measured by
 139 current well-calibrated LEO (low-earth orbiting; e.g., Aura/OMI, Ozone Monitoring Instrument, and
 140 Suomi-NPP/OMPS, National Polar-orbiting Partnership/Ozone Mapping and Profiler Suite) satellite
 141 instruments observing scenes that match in time and observing angles with those from EPIC. For albedo
 142 measurements, OMPS has a calibration accuracy of 2 %, while its wavelength dependence (precision) in
 143 the calibration is estimated to be better than 1 % (Jaross et al., 2014). The OMPS Nadir Mapper on
 144 Suomi-NPP has a 50 x 50 km² footprint in its normal operating mode with 36 cross-track views ($\pm 55^\circ$
 145 satellite view angle or strip of about $\pm 12^\circ$ equatorial longitude). It has a spectral resolution of 1 nm,
 146 which is close to EPIC's 317.5 nm and 325 nm channels FWHM, but narrower than EPIC's 340 nm and
 147 388 nm channels. To perform in-flight calibration, OMPS' albedo spectra were either interpolated (for
 148 317.5 and 325 nm channels) or convolved (at 340 and 388 nm) with each EPIC filter transmission
 149 function T_M (Fig. 1). Because the albedo spectra $A_M(\lambda)$ (Eq. 1) removes the Fraunhofer line structure
 150 contained in both the solar irradiance S_M and the reflected Earth radiance $I_M(\lambda)$, the interpolation and
 151 convolution of $A_M(\lambda)$ has better accuracy than directly using $I_M(\lambda)$. OMI on Aura has 13 x 24 km² spatial
 152 resolution and about $\pm 56^\circ$ cross-track views (a strip of ± 1300 km or $\pm 13^\circ$ equatorial longitude) with a
 153 spectral resolution of 0.42 nm. To match measurements with DSCOVR, OMI's albedo spectra were
 154 convolved with EPIC's $T_M(\lambda)$. Then, the results in every two adjacent cross-track views and four
 155 consecutive along-track scans are combined to form 50 x 50 km² footprints for comparison with EPIC
 156 measured counts/second obtained from 7 x 7 EPIC pixels.

157 EPIC raw counts/second inside each coincident footprint are preprocessed by the steps stated in
 158 a previous paragraph. Then, the counts/second average and variance in each coincident footprint are
 159 computed to obtain the EPIC albedo calibration coefficients K_M (Eq. 3). Misalignment between EPIC and
 160 OMPS or OMI footprints can result large scene noise unless uniform scenes are selected and less
 161 uniform scenes discarded. This is achieved by weighting each coincident data point with the reciprocal
 162 of the percent EPIC counts/second variance inside the coincident footprint. All of the coincident points
 163 between LEO satellites and EPIC observations occur within $\pm 40^\circ$ of the earth's equator. Selected LEO
 164 footprints have viewing angles nearly identical to EPIC's (within 1° in backscatter angle and 2° degrees in
 165 solar zenith angle). EPIC's backscatter angle varies with latitude and longitude by less than 0.25° , since
 166 the angular size of the earth varies from 0.45° to 0.53° to 0.45° every 6 months depending on the
 167 location of DSCOVR in its orbit (an irregular Lissajous orbit about L-1 that is tilted relative to the ecliptic
 168 plane and perturbed by the Earth's moon). The orbit varies from 4° to 15° away from the Earth-Sun line.
 169 These small differences in observing geometry are corrected in the atmospheric radiative transfer model
 170 calculations $\alpha(\lambda)$ (Eq. 4), resulting in corrections less than 2 %. EPIC albedo calibration coefficients are
 171 derived from Eqs. 3 and 4.

$$K_M = \frac{A_M(OMPS) \{ \alpha_M(EPIC) / \alpha_M(OMPS) \}}{C_M(EPIC) D_E^2} \quad (3)$$

$$\alpha_M = \int \alpha(\lambda) S(\lambda) T_M(\lambda) d\lambda / \int S(\lambda) T_M(\lambda) d\lambda \quad (4)$$

172

173 where

174 M is the EPIC channel number, M=1,2,3,4

175 $A_M(\text{OMPS})$ = OMPS albedo measurement in the EPIC channel-M wavelength band

176 $\alpha_M(\text{EPIC})$ and $\alpha_M(\text{OMPS})$ are computed albedo values for EPIC and OMPS coincident geometry,

177 $C_M(\text{EPIC})$ is the average count rate over the pixels matching OMPS,

178 D_E is the sun-earth distance in AU.

179 $\alpha(\lambda)$ is the computed high resolution normalized radiance spectrum,

180 $S(\lambda)$ is the referenced high resolution solar irradiance spectrum,

181 $T_M(\lambda)$ is the EPIC filter transmission profile or the OMPS slit function.

182

183 All of the coincidence points with LEO satellite instruments were measured using the area of
184 the EPIC CCD within 600 pixels of its center. There are about 15000 coincidence data points accumulated
185 by the end of 2016. Because of the large number of data points, statistical averaging errors are small.
186 An atmospheric radiative transfer model, RTM, takes total column ozone and surface reflectivity from
187 LEO retrievals to obtain both $\alpha_M(\text{EPIC})$ and $\alpha_M(\text{LEO})$. Although uncertainties in the RTM can propagate
188 into the computed albedos, the resulting uncertainties in $\alpha_M(\text{EPIC})$ and $\alpha_M(\text{LEO})$ are approximately
189 identical, and approximately cancel in Eq. 3. The resulting EPIC albedo calibration uncertainty is mostly
190 inherited from the OMPS albedo calibration uncertainty, which has an accuracy of 2 % and a precision of
191 1 % in relative (wavelength dependent) values. For the UV channels, the calibration factors K_M are not
192 constants, but are slowly increasing functions of time (on average 0.016 per year; see $K_M(t)$ in Fig. 2),
193 which is normalized to one on 1 January 2016). Table 2 shows the reference values of K_M multiplied by
194 π .

195 Using Tables 1, 2, and Fig. 2, EPIC albedo measurements are derived with

196

$$A_M(\text{EPIC}) = K_M C_M(\text{EPIC}) D_E^2 \quad (1-5)$$

197

198

199 Note that the factor D_E^2 for solar irradiance at 1 AU is contained in the albedo calibration
200 coefficient K_M . Since solar activity changes (e.g., 27.5 day cycle) are negligible for EPIC UV channel
201 wavelengths, daily solar irradiance changes are only adjusted with the sun-earth distance D_E . Users of
202 EPIC data may also be interested in radiance measurements. The radiance calibration coefficients can
203 be derived with Eq. 6,

204

$$E_M = K_M S_M \quad (6)$$

205

206

207 and the radiance measurements can be obtained with Eq. 7.

208

$$I_M(\text{EPIC}) = E_M C_M(\text{EPIC}) \quad (7)$$

209

210

211 The uncertainty in the radiance calibration can increase significantly due to errors in estimating
 212 the absolute solar irradiance. Uncertainty in estimated S_M for EPIC UV channels in Table 1 is about 3 %.

213 **1.3 Ozone Algorithm**

214

215 Once the albedo calibration factors are applied to EPIC's measured counts/second, the
 216 calculated albedos can be combined to retrieve total column ozone (TCO), Lambert Equivalent
 217 Reflectivity (LER), and aerosol index (AI). The TOA directional albedo calculation uses the TOMRAD
 218 radiative transfer calculation code, which has a spherical geometry correction for large solar zenith
 219 angles (SZA) and satellite looking angles (SLA) (Caudill et al., 1997). The calculation uses the same
 220 climatological ozone profiles used in OMI retrievals, altitude weighted average effective ozone
 221 temperatures, ground reflectivities, terrain height, and climatological cloud heights. Spectrally resolved
 222 O_3 absorption cross sections are from Brion et al., (1993, 1998); Daumont et al., (1992); and Malicet et
 223 al., (1995). The resulting spectra are convolved with the EPIC filter transmission functions (Fig. 1) and
 224 with the reference solar irradiance spectra (see Eq. 4).

225
 226 The resulting computed α_M (Eq. 4) are compiled into a finely stepped look-up table as functions
 227 of ozone profiles and solar-view angles. EPIC ozone retrieval uses the 388 nm channel for computing
 228 the surface reflectivity with a formula similar (except for choice of wavelengths) to that used in cloud
 229 reflectivity studies (Herman et al., 2009). Then, the retrieval is based on two ozone absorption channels,
 230 317.5 nm and 340 nm for low optical depth conditions, or 325 nm and 340 nm for high optical depth
 231 conditions, together with the 388 nm measurement to form triplet equations. The ozone retrieval
 232 algorithm assumes a linear wavelength dependence in the surface reflectivity (Eq. 8),

$$R_\lambda = R_{\lambda_0} + b(\lambda - \lambda_0) \quad (8)$$

233
 234 where λ_0 is given wavelength 388 nm. The total column ozone (TCO) is given by Eq. 9,

$$\Omega = \Omega_0 + \frac{\Delta N_{\lambda_1} \frac{\partial N_{\lambda_2}}{\partial R} (\lambda_2 - \lambda_0) - \Delta N_{\lambda_2} \frac{\partial N_{\lambda_1}}{\partial R} (\lambda_1 - \lambda_0)}{\frac{\partial N_{\lambda_1}}{\partial \Omega} \frac{\partial N_{\lambda_2}}{\partial R} (\lambda_2 - \lambda_0) - \frac{\partial N_{\lambda_2}}{\partial \Omega} \frac{\partial N_{\lambda_1}}{\partial R} (\lambda_1 - \lambda_0)} \quad (9)$$

235 where

236 Ω_0 is an initial climatology estimate of TCO or TCO from previous step in the iteration,
 237 λ_1 and λ_2 are the selected ozone absorption wavelengths,
 238 N_λ is the N-value defined as logarithm of the albedo values by Eq. 10,

$$N_\lambda = -100 \log_{10}\{I_\lambda / (S_\lambda / D^2)\} \quad (10)$$

240
 241 and

242 ΔN_λ is the N-value residue (difference between the measured N-value and the computed N-value),

243 $\frac{\partial N_{\lambda x}}{\partial Z}$ = measurement sensitivity with respect to the total column ozone, $Z = \Omega$, or the surface
244 reflectivity, $Z = R$, for wavelengths λ_1 or λ_2 .

245 If one assumes the sensitivities to the surface reflectivity, $\partial N_{\lambda}/\partial R$ are wavelength independent,
246 Eq. 5 for the triplet algorithm is similar to the Version 8 TOMS algorithm (Rodriguez et al., 2003).

247 Since the algorithm for ozone (Eqs. 8 to 10) requires the use of two or more wavelength
248 channels, the measured counts/second for each channel must be geolocated on a common latitude x
249 longitude grid that is accurate to 0.25 of a single pixel size. When projected on the 3-D Earth, the
250 sampling size is about 8 km at nadir and effectively increases to 10 km when EPIC's point spread
251 function is applied. The result for 2 x 2 pixel averaging is a spatial resolution at nadir of about 18 km,
252 which gets larger as the secant of the SLA from the nadir point. SLA is measured relative to the normal
253 to the Earth's surface, and is 0° at nadir and almost 90° at the Earth's sunlit terminator. The radiative
254 transfer spherical geometry correction is accurate to about 80° in SZA and SLA, which means that
255 retrieved ozone values near the Earth's terminator are not accurate.

256 **2 Natural Color Images**

257 A typical eye response color image view of the Earth, obtained by a weighted combination of
258 the geolocated red, green, and blue wavelength channels, is shown in Fig. 2. To produce RGB images
259 adjusted to the human eye response, the algorithm used is a derivative of the International Commission
260 on Illumination (CIE) process for estimating tristimulus values from calibrated instruments (Wyszecki
261 and Stiles, 1982; Broadbent, 2004; Gardner, 2007; Bodrogi and Khanh, 2012). Obtaining eye response
262 images for EPIC's narrow band filters (Table 1) was improved by customization of the algorithm to use
263 additional channels than just the 443, 551, and 680 nm blue, green, and red channels.

264 Because the blue 443 nm channel is not spatially averaged onboard the spacecraft, the color
265 images have a maximum resolution of about 10 km at nadir determined by looking at the discernable
266 width of the Nile and Amazon Rivers. The color images also give an indication of the quality of the
267 geolocation. Errors in geolocation would appear as pink edges at the cloud boundaries, which are not
268 present in the images in Figs. 3 or in the complete image collection on <http://epic.gsfc.nasa.gov/>.

269 Even with accurate geolocation, about 0.25 pixels (2 km), between the 4 UV channels, there is
270 some noise introduced into ozone retrievals by small cloud edge location errors when transferring all of
271 the native data to a common latitude and longitude grid. Ozone retrievals over almost cloud-free
272 scenes, such as over the Saharan desert or clear-sky portions of the oceans, show much less noise than
273 those with partial cloud cover. Since the pixel-to-pixel noise caused by misaligned cloud edges is almost
274 random, spatial averaging to about $50 \times 50 \text{ km}^2$ (similar to TOMS and OMPS, but coarser than OMI spatial
275 resolution) reduces the effect of apparent noise from cloud edges. The following sections use 25×25
276 km^2 spatial averaging (3 x 3 CCD pixels), which has more spatial details and some cloud-edge noise
277 (noise < 3 %).

278

279 **3 Examples of EPIC Ozone and Reflectivity**

280 A matched pair of images for ozone and scene reflectivity LER (17 April 2016) are shown in Fig. 4
281 with a maximum resolution of 18 km, since all UV channels involved in the ozone retrieval are
282 downlinked from the spacecraft at a resolution of 2 x 2 onboard averaged pixels. Note that the reduced
283 resolution hdf5 data files stored on the ground are in their original sampling density (2048 x 2048), but
284 have reduced spatial resolution. In Fig. 4, the entire data image for ozone and the LER scene reflectivity
285 are all at a common Universal Time (00:36 UTC or 12:36 local time at the center of the image) and
286 encompasses local times from sunrise (west) to sunset (east) with all images rotated so that north is up.
287 In the LER scene, a large east-west belt of clouds are visible near the equator, as are cloud plumes
288 descending from the Arctic. The major cloud patterns change slowly, but show major seasonal changes.
289 Figure 5 shows six additional scenes from the same day, 17 April 2016, with large cloud features
290 associated with the Arctic region, an equatorial cloud band, and large cloud structures over the Antarctic
291 Ocean. Figure 6 shows reflectivity measurements for 23 November 2015 with cloud features common in
292 the Southern Hemisphere SH. The cloud band extending toward the Antarctic region from Argentina's
293 Salado River is an example of a persistent feature that appears frequently throughout the year. In a
294 later section, the amounts of retrieved ozone and cloud reflectivity $0 < R_c < 1$ are used to estimate the
295 amount of UV radiation reaching the earth's surface over snow/ice free scenes.

296 The Arctic and Antarctic ice sheets are visible after their spring equinox times, and especially in
297 their respective late spring and summer images when the Earth's poles are tilted toward L-1 (Figs. 5 and
298 6). In the color and LER images, clouds over ice are not readily visible because of the very high ice
299 reflectivity providing little or no contrast with 388 nm cloud reflectivity. It is possible to obtain
300 information about clouds over ice from the O₂ A-band channel at 764 nm (Fig. 7), which differentiates
301 between reflecting surfaces that are at different altitudes because of oxygen absorption in the
302 atmosphere. In this image, the bright white clouds (less atmospheric O₂ absorption) are at higher
303 altitudes than the grey clouds, which are all higher than the ice surfaces. A quantitative analysis of cloud
304 height and cloud-caused reduction in solar irradiance reaching the ice surface will be the subject of a
305 future paper.

306 **4 Validation of EPIC Ozone Retrieval**

307 EPIC retrieved ozone can be validated by comparison with other ozone measuring satellite data
308 (e.g., OMI, and OMPS) and by comparison with well-calibrated ground-based instruments.

309 While EPIC observes from sunrise to sunset in every image, there are only 6 to 8 useful
310 coincidences per 24 hours with a specified ground site separated by either 68 minutes (NH summer) or
311 110 minutes (NH winter). Coincidences at high SZA $> 75^\circ$ are increasingly inaccurate for both satellite
312 and ground-based retrievals. This problem is compounded for EPIC, since high SZA also implies high SLA,
313 which increases the spherical geometry correction error. Ozone absorption and Rayleigh scattering at
314 high SZA also prevents 317.5 nm radiances from reaching into the lower troposphere and to the surface,
315 which is partially mitigated by having the retrieval algorithm automatically switch from 317.5 nm to 325
316 nm at high optical depths (usually high SZA).

317 A comparison of EPIC retrieved TCO with those determined by a Pandora spectrometer
318 instrument (#034) located at Boulder, Colorado is shown in Fig. 8. This Pandora was selected because it
319 has been extensively compared to a well calibrated Dobson spectroradiometer and to OMI and OMPS
320 ozone overpass data (Herman et al., 2015). The Pandora data are matched in location and time t_0 to the
321 EPIC UTC when Boulder, Colorado is in view (several times per 24 hours). Pandora ozone is averaged
322 over $t_0 \pm 12$ minutes. EPIC data are limited to distances within 50 km of Boulder, Colorado. Figure 8
323 shows that EPIC and Pandora ozone amounts track each other closely during 2015 and 2016. The 2015-
324 2016 average agreement is 2.7 ± 4.9 %. There is a period in the winter of 2016 where the Pandora data
325 quality was degraded by the presence of heavy cloud cover and in February by a mechanical problem
326 with the Pandora sun tracker.

327 The OMI and OMPS satellites are polar orbiting with an equator crossing time of about 13:30
328 hours local time measuring in a narrow strip on either side of the orbital track. While it is possible to
329 compare EPIC ozone with low earth orbit satellite data, a more complete comparison can be made with
330 the assimilated ozone product from MERRA-2, the Modern-Era Retrospective Analysis for Research and
331 Applications, (Rotman et al., 2002) version 2 (MERRA-2, Molod et al., 2014). MERRA-2 ozone is based on
332 Microwave Limb Sounder (MLS) and total column ozone from the Ozone Monitoring Instrument OMI on
333 NASA's EOS *Aura* satellite. The advantage of using MERRA-2 is that the ozone field is synoptic and can be
334 directly compared with EPIC for the same UTC (Fig. 9) over the same sunlit globe as seen by EPIC. The ozone
335 structures seen by EPIC are all present in the MERRA-2 independent assimilation, even though there is an
336 average offset of about 10 DU (3 %). The disagreement with EPIC is similar to the offset of MERRA-2 with
337 other satellite data (Wargan et al., 2017). A close look at the ozone maps in Fig. 9 shows overall agreement
338 with most features including the small region of elevated O_3 over the central US. There are differences, such
339 as the higher amount of O_3 measured by EPIC over Brazil on 23 November and the structure at $15^\circ N$ in the
340 transition from equatorial O_3 values to mid-latitude values (dark blue to light blue).

341 **5.0 Synoptic Variation of Ozone (SVO) from Sunrise to Sunset**

342 Most LEO satellite views of ozone are at almost fixed local time based on the equator crossing
343 local solar time (13.5 ± 0.8 hours side scanning) with approximately 20 minutes local time variation from
344 the equator to the pole. Longitudinal coverage is obtained by piecing together North-South strips
345 obtained about 90 minutes apart. Variation that occurs on a scale less than 90 minutes cannot be seen
346 from a polar orbiting LEO satellite, nor can variation from different local times of the day. EPIC observes
347 from close to sunrise and sunset with local solar noon near the center of the data set as shown in Fig.
348 10. The exact position of noon in the EPIC images depends on the location of EPIC in its orbit relative to
349 the Earth-Sun line. The longitude resolution is approximately 0.25° at the center of the FOV, which
350 corresponds to a time resolution of about 1 minute. The resolution decreases as the secant of the angle
351 from the center (e.g., 2 minutes or 0.5° at 60° from the center). A limitation in the EPIC observations
352 occurs at high SZA and high SLA. As can be seen in Fig. 10, ozone values near the morning terminator are
353 probably too low compared to the middle longitude values. These retrieval errors are partly caused by
354 the effects of spherical geometry that are not properly represented in the TOMRAD radiative transfer
355 calculations.

356 The view of the EPIC instrument from sunrise to sunset at fixed UTC is not the diurnal variation
357 that an instrument on the ground would see from sunrise to sunset. For the ground-based Pandora
358 instrument, the observed changes throughout the day from sunrise to sunset are at varying UTC every
359 80 seconds. Compared to the ground-based viewpoint, EPIC obtains data for a fixed geographic location
360 every 68 minutes UTC in NH daytime summer and every 110 minutes in NH daytime winter.

361 **5.1 Southern Hemisphere SH Late Spring 23 November 2015 :**

362 To illustrate the SH synoptic change in ozone, Figs. 10 and 11 show the diurnal (longitudinal)
363 variation of ozone centered on the South American continent on 23 November 2015 at 16:20 UTC. The
364 local time varies from early morning (06:20, -150° longitude) to late-afternoon (16:20, 0° longitude). At
365 high southern latitudes, 60°S and 70°S, the late spring (23 November) residue of 2015 Antarctic ozone
366 hole is clearly visible in the ozone map image (Fig. 10). Figure 11 shows details of the ozone amounts in
367 specified latitude bands ($\pm 0.125^\circ$ wide) in the Southern Hemisphere sampled every 5° degrees from 0°
368 to 70°S. Solar zenith angles are limited to the range $\pm 70^\circ$ to avoid high latitudes and longitudes near
369 sunrise or sunset where spherical geometry effects become important. This particular example (Fig. 11)
370 is from one image centered over South America (Fig. 10). For 23 November there are 15 more
371 overlapping images covering the entire 360° of longitude that could be combined to produce a complete
372 composite global map of ozone at 15 different UTCs. In the NH summer there would be 22 images per
373 day. A composite ozone map of this kind would no longer be synoptic, since overlapping data are
374 averaged, but would now be similar to the joined data strips from OMI or OMPS.

375 Figure 11 contains the data points from a $0.25^\circ \times 0.25^\circ$ average within each 5° latitude band L
376 shown as light grey dots. The dark lines are a Lowess(0.05) fit (locally weighted least squares fit to 5 % of
377 the data, (Cleveland, 1981)), which corresponds to approximately a 30 minute time average (7.5°
378 Longitude). The largest apparent scatter from the Lowess fit occurs at $L = 0.125^\circ$ S, which amounts to a
379 longitudinal standard deviation from the mean of ± 4 DU or ± 1.5 %. The equatorial bands (0° S to 20° S)
380 shows considerable longitudinal change (10–20 % from $L = 0$ – 40° S rising to 75 % at $L = 70^\circ$ S,
381 approximately as $TCO = 16.063 + 0.56L + 0.02L^2$). Most of the observed changes are dynamically driven,
382 since the photochemistry involved in the stratosphere (20 - 25 km altitude) is too slow to produce such
383 large changes with changing SZA. Southward of 45° S, the effects of the remaining ozone hole depletion
384 (dark blue in Fig. 10), which is still present in November, appear at -50° longitude as indicated in Fig. 11.

385 **5.2 Northern Hemisphere NH Summer Solstice 21 June 2016:**

386 An example is provided for the ozone retrievals obtained on 21 June 2016 at 18:41 UTC that is
387 approximately centered over North America (Fig. 12). Since this is Northern Hemisphere summer
388 solstice, corresponding to the sun being nearly overhead at 23° N, the latitude range available for
389 retrieving ozone extends over the North Pole. Figure 13 contains ozone retrievals in 0.25° wide latitude
390 bands similar to Fig. 11. Unlike the SH 23 November 2015 example, there is only moderate longitudinal
391 (diurnal) variability in ozone amount for latitudes between 0° and 15° N. However, there is a clear wave
392 structure in the 20° N to 25° N bands with a periodicity of approximately 35° longitude (2.3 hours) and
393 again in the 40° N to 60° N bands that are not obvious in the global map (Fig. 12).

394 The dynamical effects on ozone in the NH mid-latitudes are quite different than their
395 counterparts in the SH, where the NH mid-latitude behavior (30°N – 35°N) is clearly separated from
396 equatorial and high latitude bands with an increase in ozone amount from about 280 DU to about 350
397 DU, which is larger than a similar increase in the SH. There is an ozone periodicity of approximately 38°
398 longitude (2.5 hours) at 30°N – 35°N midday and a longer longitudinal period 73° (4.9 hours) in the
399 morning. At higher latitudes, 35°N – 55°N , the variability is more pronounced with an approximate
400 period of 55° (3.6 hours). In the bands from 55°N – 70°N the variability is reduced and the ozone amount
401 falls from mid-latitude values of about 350 DU to below 300 DU. The wave structure varies throughout
402 the year in both hemispheres.

403 **5.3 Northern and Southern Hemisphere 17 April 2016 18:35 UTC**

404 Figure 5-5 shows the ozone retrieval for the sunlit globe on 17 April 2016 at 18:36 UTC about 1
405 month from the March equinox including large plumes of elevated ozone amounts (450 DU) extending
406 from high latitudes into mid-latitudes where the usual ozone amount is about 350 DU. For the SH (Fig. 5-
407 5), polar ozone variability (280-320 DU) is relatively small compared to November 23 (Fig. 10). There is
408 wave structure (Fig. 15) between 30°S and 40°S with a periodicity of about 4 hours (60° longitude) (see
409 also Schoeberl and Kreuger, 1983). The dip in O_3 amount at 77°W to 67°W and 10°S to 25°S
410 corresponds to the Andes Mountains in Peru, Bolivia, and Chile. While the SZA range is limited to $\pm 70^{\circ}$,
411 the SLA reaches more than 80° at low latitudes for longitudes between 40°S and 20°S introducing
412 spherical geometry correction errors that increase towards sunset near 20°W . The errors appear as
413 apparent increases in O_3 amount. At higher latitudes, the SLA is in the middle 70° s when the SZA is 70° .
414 The high SLA error is present in both hemispheres for observations near equinox.

415 The NH shows little variability in the equatorial region (0 – 25°N) with a mean value of about 260
416 DU (Fig. 16). The SLA error is present for latitudes between 0 and 15°N and 0 and 15°S that appears as
417 an elevated ozone amount at longitudes east of 50°W . Mid-latitudes (30°N to 40°N) show a wave
418 structure that is approximately 37° apart (2.5 hours) at 35°N . A similar structure occurs in the SH with a
419 period of about 4.5 hours. There is an ozone maximum (red area in Fig. 14 about 450 DU) near 140°W
420 extending from 60°N to 35°N , very high ozone amounts in the Arctic region, and a high ozone patch
421 over the central US (35°N to 45°N and 104°W) peaking at 420 DU (40°N and 104°W), which probably
422 corresponds to a region of high atmospheric pressure.

423 **6.0 Estimating Erythemal Irradiance at the Earth's Surface**

424 The unique observing geometry of DSCOVR/EPIC permit the use of synoptic ozone and cloud
425 reflectivity data to be used to compute the diurnal variation of UV irradiance from sunrise to sunset for
426 any point on the illuminated earth observed by EPIC. Previous calculations from satellite data used
427 cloud cover and ozone from 13:30 and assumed it applied to local noon. The assumption is usually
428 adequate for slowly varying ozone, but not for estimating the effects of more rapidly varying cloud
429 cover. The following paragraphs discuss the calculation of erythemal irradiance, a spectrally weighted
430 mixture of UV wavelengths used as a measure of skin reddening and potential sunburn from exposure to
431 sunlight.

432 Erythemal irradiance $E_0(\text{SZA } \theta, C_T)$ at the earth sea level (watts/m²) is defined in terms of a
 433 wavelength dependent weighted integral over a specified weighting function $A(\lambda)$ times the incident
 434 solar irradiance $I(\lambda, \theta, \Omega, C_T)$ (Watts/m²) (Eq. 11) at the Earth's sea level. The erythemal weighting function
 435 $\text{Log}_{10}(A_{\text{ERY}}(\lambda))$ is given by the standard Erythemal fitting function shown in Eq. 12 (McKinley and Diffey,
 436 1987). Tables of radiative transfer solutions for $D_E = 1$ AU are generated for a range of sza ($0 < \theta < 90^\circ$),
 437 for ozone amounts $100 < \Omega < 600$ DU, and terrain heights $0 < Z < 5$ km using the TUV DISORT radiative
 438 transfer model as described in Herman (2010) for erythemal and other action spectra (e.g., plant
 439 growth, vitamin D production, cataracts, etc.).

$$E_0(\theta, \Omega, C_T) = \int_{250}^{400} I(\lambda, \theta, \Omega, C_T) A(\lambda) d\lambda \quad (11)$$

$$\begin{aligned} 250 < \lambda < 298 \text{ nm} & \quad \text{Log}_{10}(A_{\text{ERY}}) = 0 & (12) \\ 298 < \lambda < 328 \text{ nm} & \quad \text{Log}_{10}(A_{\text{ERY}}) = 0.094 (298 - \lambda) \\ 328 < \lambda < 400 \text{ nm} & \quad \text{Log}_{10}(A_{\text{ERY}}) = 0.015 (139 - \lambda) \end{aligned}$$

440 Equation 11 can be accurately approximated by the power law form (Eq. 13), where $U(\theta)$ and $R(\theta)$
 441 are fitting coefficients to the radiative transfer solutions in the form of rational fractions. Rational
 442 fractions were chosen because they tend to behave better at the ends of the fitting range than
 443 comparable fitting accuracy polynomials.

$$E_0(\theta, \Omega, C_T) = U(\theta) (\Omega/200)^{-R(\theta)} C_T \quad (13)$$

$$U(\theta) \text{ or } R(\theta) = (a+c\theta^2+e\theta^4)/(1+b\theta^2+d\theta^4+f\theta^6) \quad r^2 > 0.9999 \quad (14)$$

$$C_T = (1-\text{LER})/(1-R_G) \text{ where } R_G \text{ is the reflectivity of the surface} \quad (15)$$

$$E(\theta, \Omega, Z) = E_0(\theta, \Omega) H(\theta, \Omega, z) \quad (16)$$

$$H(\theta, \Omega, Z) = 1+(0.04652 Z_{\text{km}} + 0.00496) (-0.07033 (\Omega/200) + 1.12303)G(\theta) \quad (17)$$

$$G(\theta) = g+h\theta+i\theta^2+j\theta^3+k\theta^4 \quad (18)$$

The coefficients $a, b, c, d, e, f, g, h, j,$ and k are in Tables A-1 and A-2 in the appendix

444 The E_0 solutions to the radiative transfer calculations can be accurately reproduced by a relatively
 445 simple functional form (Eqs. 13 to 15) with the coefficients given in Table A-1. These are the same
 446 coefficients given in Herman (2010) along with other biological action spectra weighting functions, $H(z, \theta)$
 447 is a function representing the increase in $E(\theta, \Omega, Z)$ with altitude per km, and C_T is the cloud transmission
 448 function (Eq. 15) estimated from the retrieved LER derived by assuming that the cloud-ground system
 449 can be approximated by a two-layer Stokes problem (elevated cloud and surface) with atmospheric
 450 effects between the cloud bottom and the surface neglected (Herman et al., 2009). r^2 is a measure of
 451 the correlation of the E_0 data points with the fitting function. Eqs. 13 to 18 are for an Earth-Sun distance
 452 of 1 AU.

453 For E_0 The fitting residual is less than $\pm 0.001 \text{ W/m}^2$ compared to the worst case when $E_0(50^\circ,$
454 $200) = 0.15 \text{ W/m}^2$ (Herman, 2010). When height effects are included $E(\theta, \Omega, Z) = E_0(\theta, \Omega) H(\theta, \Omega, Z)$,
455 where $H(\theta, \Omega, Z)$ is a fitting polynomial (Eq. 17) to the downward irradiance at 0, 1, 2, 3, 4, and 5 km
456 based on results from the radiative transfer calculation. The increase of erythemal irradiance with
457 altitude has an SZA dependence given by $G(\theta)$, which increases with θ until θ is approximately 60° , and
458 then $G(\theta)$ decreases.

459 The height dependence of $E(\theta, \Omega, Z)$ is similar to that derived by Chubarova et al. (2016) for low
460 aerosol amounts. When absorbing aerosols have a significant optical depth, Chubarova et al. (2016)
461 derived a multiplicative correction term to $E(\theta, \Omega, Z)$ for a wide variety of conditions.

462
463 When Eq. 13 is applied to the ozone and LER data described in previous sections, the global
464 erythemal irradiance at the ground can be obtained after correction for the Earth-Sun distance D_E in a
465 manner similar to Eq. 1, where D_E in AU can be approximated by (Eq. 19),
466

$$D_E = 1 - 0.01672 \cos(360 (\text{day_of_year} - 4)/365.25) \quad (19)$$

467
468 An example of $E(\theta, \Omega, Z)$ is shown in Fig. 17 for 17 April 2016 at 18:35 UTC. Local noon is near
469 the center of the image with sunrise to the left (west) and sunset to the right (east). For this date, the
470 sun is overhead just north of the equator producing very high values of Erythemal irradiance $E(\theta, \Omega, Z)$
471 corresponding to a UV index, UVI, of 13 at sea level in the Pacific Ocean (UVI = 40 $E(\theta, \Omega, Z)$). The UVI
472 scale was designed for sea level mid latitudes ranging from 0 to 10 to provide public health warnings
473 (e.g. for UVI = 8). Somewhat higher values are seen in the Sierra Nevada Mountains in Mexico near
474 20°N . This particular day is relatively cloud free over most of South America except for clouds over
475 southern Brazil extending into Paraguay and other small patches of clouds. For the erythemal irradiance,
476 the presence of clouds reduces the amount of UV reaching the ground (blue color with a UV index of
477 less than 4).

478
479 The increase with altitude is much more pronounced during the summer months over the Andes
480 Mountains reaching above 4 km (over 13,000 feet). Figures 18 and 19 show the large increases with
481 altitude over the Andes Mountains for 23 November 2015, with the sun nearly overhead at 20°S
482 latitude. Here the UV index ranges from 16 to 18, which agrees with previous ground-based
483 measurements in this region (Cede et al., 2002). Any significant unprotected exposure to these levels of
484 UV would lead to severe sunburn and eye damage. On a completely clear day the UV index would be
485 even higher than 18. Figure 19 is a longitudinal slice through the UV data in Fig. 18 at 20°S . The figure
486 shows the longitudinal variation $E(\theta, \Omega, Z)$ as a function of local time, the effect of light clouds on the
487 eastern side of the Andes Mountains, and the sharp reduction at 50°W .

488
489 Figure 20 shows the erythemal irradiance computed for 21 June 2016 centered over the US and
490 Central America. The sun is overhead at 23.3°N latitude. In the clear regions not covered with light
491 clouds, the UV index reaches about 12 extending from an area in the Pacific Ocean at 15°N up into the
492 US mid-west, Rocky Mountains, Utah and New Mexico. The eastern US has a lower UV index of about 8.

493 The extended scale of this map (UVI = 0 to 20) is too coarse to see the variation with latitude on the east
494 coast.

495
496 Similarly, Fig. 21 shows high values of Erythemal irradiance in the Himalayan Mountains on June
497 21, 2016 with peak UV index of about 15 even in the presence of partial cloud cover that reflects a
498 portion of the incident solar flux back to space. The effect of cloud cover can be seen in Fig. 22, which is
499 a longitudinal slice through the irradiance values associated with the latitude at 32°N. In the absence of
500 clouds, the peak value of the UV index would be close to 20. Even with cloud cover, the UV index
501 reached 15, which is twice the value of a typical cloudless summer case in the US at comparable
502 latitude.

503 **7.0 Summary**

504 The DSCOVR/EPIC 10-filter Spectroradiometer (317.5 to 780 nm) makes measurements of the
505 the rotating sunlit face of the earth from the Lagrange-1 point located 1.5×10^6 km from the earth with a
506 maximum resolution of $10 \times 10 \text{ km}^2$ for 443 nm at the sub-satellite point. The other 9 channels have $18 \times$
507 18 km^2 resolution. The key difference between EPIC and LEO satellites is EPIC's ability to measure the
508 whole sunlit earth (sunrise to sunset) at the same UTC (synoptic measurements) every 68 or 110
509 minutes depending on the season at the Wallops Island, Virginia data receiving station. EPIC ozone
510 retrievals have been compared successfully to both ground-based Pandora spectrometer instruments
511 and to the MERRA-2 satellite data assimilation model for the same UTC observed by EPIC. EPIC's
512 synoptic measurements insure that the ozone amounts, cloud reflectivity, and aerosol amounts that are
513 used to estimate UV irradiance are the proper values for each time of the day. EPIC has been making
514 measurements since June 15, 2015 with no evidence of significant degradation relative to LEO satellites
515 observing the same scene at the same angles. EPIC has obtained ozone and reflectivity data multiple
516 times per 24 hours for over two years that can be used to more accurately estimate the health effects
517 from continuous or periodic exposure during any day to UV radiation reaching the ground including the
518 effects of cloud cover and altitude.

519

520

521

522

523

524

525

526

527 **Appendix**

528 Figure A1 illustrates the orbit of the DSCOVER spacecraft following the earth in its orbit about the
 529 sun.

530

Table A-1 Coefficients $R(\theta)$ and scaling coefficient $U(\theta)$ for $0 < \theta < 80^\circ$
 and $100 < \Omega < 600$ DU for $E(\Omega, \theta) = U(\theta) (\Omega/200)^{-R(\theta)}$ ($1.0E10 = 1.0 \times 10^{10}$)

$$U(\theta) \text{ or } R(\theta) = (a+c\theta^2+e\theta^4)/(1+b\theta^2+d\theta^4+f\theta^6) \quad r^2 > 0.9999$$

Action Spectra	$U(\theta)$ (watts/m ²)	$R(\theta)$
CIE Erythemat U_{ERY} & R_{ERY}	a= 0.4703918683355716 b= 0.0001485533527344676 c= -0.0001188976502179551 d= 1.915618238117361E-08 e= 7.693069873238405E-09 f= 1.633190561844982E-12	a= 1.203020609002682 b= -0.0001035585455444773 c= -0.00013250509260352 d= 4.953161533805639E-09 e= 1.897253186594168E-09 f= 0.0

Table A-2 Solar Zenith angle function $G(\theta)$ used in Eq. 18

$$G(\theta) = g+h\theta+i\theta^2+j\theta^3+k\theta^4$$

g= 0.9996074048174048	j= 1.412462444962443E-06
h= 0.0001453776871276851	k= -2.037907925407924E-08
i= 2.806514180264192E-05	

531

532

533

534

535

536

537

538

539

540

541 **References**

542 Bodrogi, Peter and Tran Quoc Khanh, *Illumination, Color and Imaging: Evaluation and Optimization of*
543 *Visual Displays*, Wiley-VCH Verlag GmbH & Co. KGaA. 2012.

544
545 Brion, J., Chakir, A., Daumont, D., Malicet, J., and Parisse, C.: High-resolution laboratory absorption cross
546 section of O₃. Temperature effect, *Chem. Phys. Lett.*, 213, 610–612, 1993.

547
548 Brion, J., Chakir, A., Charbonnier, J., Daumont, D., Parisse, C., and Malicet, J.: Absorption spectra
549 measurements for the ozone molecule in the 350–830 nm region, *J. Atmos. Chem.*, 30, 291–299, 1998.

550 Broadbent, Arthur D. A critical review of the development of the CIE1931 RGB color-matching functions,
551 *Color Research & Applications*. 29 (4): 267–272. doi:10.1002/col.20020, 2004

552
553 Cariolle, D., and M. Déqué (1986), Southern hemisphere medium-scale waves and total ozone
554 disturbances in a spectral general circulation model, *J. Geophys. Res.*, 91(D10), 10825–10846,
555 doi:[10.1029/JD091iD10p10825](https://doi.org/10.1029/JD091iD10p10825) 1986.

556
557 Caudill, T.R., D.E. Flittner, B.M. Herman, O. Torres, and R.D. McPeters, Evaluation of the pseudo-
558 spherical approximation for backscattered ultraviolet radiances and ozone retrieval, *J. Geophys. Res.*,
559 102, 3881-3890, 1997.

560
561 Cede, A., E. Luccini, R.D. Piacentini, L. Nuñez, and M. Blumthaler, Monitoring of Erythemal Irradiance in
562 the Argentine Ultraviolet Network, *J. Geophys. Res.*, 107 (D13), 10.1029/2001JD001206, 2002.

563
564 Cleveland, William S., LOWESS: A program for smoothing scatterplots by robust locally weighted
565 regression. *The American Statistician*. 35 (1): 54. [JSTOR 2683591](https://www.jstor.org/stable/2683591). doi:[10.2307/2683591](https://doi.org/10.2307/2683591), 1981.

566
567 Daumont, D., Brion, J., Charbonnier, J., and Malicet, J.: Ozone UV spectroscopy I: Absorption cross-
568 sections at room temperature, *J. Atmos. Chem.*, 15, 145–155, 1992.

569
570 Dobber, M., Voors, R., Dirksen, R., Kleipool, Q., and Levelt, P., 2008, The high resolution solar reference
571 spectrum between 250 and 550 nm and its application to measurements with the Ozone Monitoring
572 Instrument. *Solar Phys.*, 249, 281–291, 2008.

573
574 Gardner, James L., Comparison of Calibration Methods for Tristimulus Colorimeters, Journal of Research
575 of the National Institutes of Standards and Technology. **112** (3): 129–138. [doi:10.6028/jres.112.010](https://doi.org/10.6028/jres.112.010),
576 2007.
577
578 Herman, J.R., G. Labow, N.C. Hsu, D. Larko (2009), Changes in Cloud Cover (1998-2006) Derived From
579 Reflectivity Time Series Using SeaWiFS, N7-TOMS, EP-TOMS, SBUV-2, and OMI Radiance Data, J.
580 Geophys. Res., 114, D01201, doi:10.1029/2007JD00008, 2009.
581
582 Jaross, G., P. K. Bhartia, G. Chen, M. Kowitt, M. Haken, Z. Chen, P. Xu, J. Warner, and T.
583 Kelly (2014), OMPS Limb Profiler instrument performance assessment, J. Geophys. Res.
584 Atmos., 119, 4399–4412, doi:[10.1002/2013JD020482](https://doi.org/10.1002/2013JD020482), 2014.
585
586 Madronich, S., McKenzie, R.L., Björn, L.O. and Caldwell, M.M. (1998), "Changes in biologically active
587 ultraviolet radiation reaching the Earth's surface," *Photochem. Photobiol.* **46**, 5-19, 1998.
588
589 Malicet, J., Daumont, D., Charbonnier, J., Chakir, C., Parisse, A., and Brion, J.: Ozone UV Spectroscopy.II:
590 Absorption cross sections and temperature dependence, J. Atmos. Chem., 21, 263–273, 1995.
591
592 Marshak, A., T., Várnai, and A., Kostinski (2017), Terrestrial glint seen from deep space: Oriented ice
593 crystals detected from the Lagrangian point, Geophys. Res. Lett., 44, doi:10.1002/2017GL073248, 2017.
594
595 McKinley, A. F., and B. L. Diffey (1987), A reference action spectrum for ultraviolet induced erythema in
596 human skin, in Human Exposure to Ultraviolet Radiation: Risks and Regulations, edited by W. R.
597 Passchier and B. F. M. Bosnjakovic, pp. 83 – 87, Elsevier, Amsterdam, 1987.
598
599 Molod, A., Takacs, L., Suarez, M., and Bacmeister, J., 2014: Development of the GEOS-5 atmospheric
600 general circulation model: evolution from MERRA to MERRA-2, Geosci. Model Dev. Discuss., 7, 7575-
601 7617, doi:10.5194/gmdd-7-7575-2014.
602
603 Rodriguez, Juan V., ; Colin J. Seftor ; Charles G. Wellemeyer and Kelly Chance (2003), "An overview of the
604 nadir sensor and algorithms for the NPOESS ozone mapping and profiler suite (OMPS)", Proc. SPIE 4891,

605 Optical Remote Sensing of the Atmosphere and Clouds III, 65 (April 9, 2003); doi:10.1117/12.467525;
606 <http://dx.doi.org/10.1117/12.467525>, 2003.

607
608 Rotman, D.A., J.R. Tannahill, D.E. Kinnison, P.S. Connell, D. Bergmann, D. Proctor, J.M. Rodriguez, S.J.
609 Lin, R.B. Rood, M.J. Prather, P.J. Rasch, D.B. Considine, R. Ramaroson, S.R. Kawa (2001), The Global
610 Modeling Initiative assessment model: Model description, integration and testing of the transport shell,
611 J. Geophys. Res., 106, 1669-1691, 2001.

612
613 Schoeberl, M.A. and A. J. Krueger, Medium Scale Disturbances in Total Ozone During Southern
614 Hemisphere Summer," *Bull. Amer. Met. Soc.*, 1358, 1983.

615
616 Tzortziou, M., J. R. Herman, A. Cede, and N. Abuhassan (2012), High precision, absolute total column
617 ozone measurements from the Pandora spectrometer system: Comparisons with data from a Brewer
618 double monochromator and Aura OMI, J. Geophys. Res., 117, D16303, doi:10.1029/2012JD017814,
619 2012.

620
621 Wargan, Krzysztof, Gordon Labow and Stacey Frith, Steven Pawson, Nathaniel Livesey, Gary Partyka,
622 Journal of Climate 30, DOI: <http://dx.doi.org/10.1175/JCLI-D-16-0699.1>, 2017.

623
624 Wyszecki, Gunther and W. S. Stiles, Color Science: Concepts and Methods, Quantitative Data and
625 Formulae, 2nd Edition, ISBN: 978-0-471-39918-6, John Wiley and Sons, 1982.

626

627 **Tables**

Table 1 Exposure Times for viewing the Earth and Full Moon (Earth side view)

Wavelength	Earth Exposure (ms)	Full Moon Exposure(ms)	Filter Width (nm FWHM)
317.5	654	2500	1
325	442	500	1
340	67	92	2.7
388	87	95	2.6
443	28	100	2.6
551	22	70	3
680	33	105	1.6
688	75	224	0.84
764	101	250	1.0
779.5	49	180	1.8

628

629

630

631

Table 2 πK_M on 1 January 2016

Irradiance at 1 AU

632

M	λ (nm)	πK_{M0}	$S_M(\text{mW}/\text{m}^2/\text{nm})$
---	----------------	--------------	---------------------------------------

633

1	317.478	1.216E-04	819.0
---	---------	-----------	-------

634

2	325.035	1.111E-04	807.7
---	---------	-----------	-------

635

3	339.858	1.975E-05	995.8
---	---------	-----------	-------

636

4	387.923	2.685E-05	1003.
---	---------	-----------	-------

637

Table A1 Coefficients R(θ) and scaling coefficient U(θ) for $0 < \theta < 80^\circ$
 and $100 < \Omega < 600$ DU for $E(\Omega, \theta) = U(\theta) (\Omega/200)^{-R(\theta)}$ ($1.0E10 = 1.0 \times 10^{10}$)

$$U(\theta) \text{ or } R(\theta) = (a+c\theta^2+e\theta^4)/(1+b\theta^2+d\theta^4+f\theta^6) \quad r^2 > 0.9999$$

Action Spectra	U(θ) (watts/m ²)	R(θ)
CIE Erythemat U _{ERY} & R _{ERY}	a= 0.4703918683355716 b= 0.0001485533527344676 c= -0.0001188976502179551 d= 1.915618238117361E-08 e= 7.693069873238405E-09 f= 1.633190561844982E-12	a= 1.203020609002682 b= -0.0001035585455444773 c= -0.00013250509260352 d= 4.953161533805639E-09 e= 1.897253186594168E-09 f= 0.0

638

639

Table A2 Solar Zenith angle function $G(\theta)$ used in Eq. 18

$$G(\theta) = g+h\theta+i\theta^2+j\theta^3+k\theta^4$$

$$g= 0.9996074048174048$$

$$j= 1.412462444962443E-06$$

$$h= 0.0001453776871276851$$

$$k= -2.037907925407924E-08$$

$$i= 2.806514180264192E-05$$

640

641

642 **Figure Captions**

643 f01 Filter transmission functions (percent) for the 10 EPIC wavelengths

644 f02 Normalized calibration functions referenced to its value at 4 Jan 2016 when $D_E = 1$ au. Average rate
645 of increase is 0.016 per year.

646 f03 Natural Color EPIC Earth images from June 6 and December 6, 2016 showing the field of view during
647 the respective hemispheric summers. In both of these images, 6-months apart, the EPIC orbit is to the
648 west of the Earth-Sun line causing the west side of the globe (sunrise) to appear brighter than the east
649 side (sunset). Notice the bright specular reflection over Argentina, South America embedded within a
650 cloud feature. This is thought to be from ice crystals in high clouds (Marshak et al., 2017).

651 f04 EPIC retrieved ozone and LER values for April 17, 2016 at 00:36 UTC. The ozone scale is from 100 to
652 500 DU, and the LER scale is from 0 to 100 percent.

653 f05 LER at six sequential UTC 0:36, 2:24, 4:12, 6:00, 7:48, and 9:36 from 17 April 2017 showing clouds in
654 the arctic region as the earth rotates in EPIC's field of view.

655 f06 Cloud formations from 23 Nov 2015 showing cloud cover in the Southern Hemisphere and near
656 Antarctica at 6 different UTC's, 10:56, 12:44 14:32, and 16:20, 14:32, 18:09, and 19:57.

657 f07 O₂ A-band View of Antarctica on December 6, 2015 showing clouds over ice. The white bright clouds
658 are at higher altitudes than the dull grey clouds because of a combination of less oxygen absorption and
659 higher optical depth.

660 f08 Daily O₃ data for EPIC (red) and Pandora (Grey) 2015 - 2016. Left: EPIC ozone data compared to
661 Pandora retrievals at Boulder Colorado. Right: Percent difference between EPIC and Pandora.

662 f09 Comparison of EPIC total column ozone with the MERRA-2 assimilation model ozone.

663 f10 Global image of ozone field for Fig. 11 for 23 Nov 2015 at 16:20 UTC

664 f11 Longitudinal or diurnal variation of ozone for the Southern Hemisphere every 5° degrees from 0° to
665 70S° for 23 Nov 2015 at 16:20 UTC. The grey points are the individual data points in the band. The solid
666 lines are a Lowess(0.05) fit to the data points representing a solar time average from 0.6 to 0.7 hours
667 depending on latitude. The SZA is limited to ±70°. Longitude = 0 Corresponds to 16:20 local time and
668 longitude = -150 corresponds to 06:20 local time.

669 f12 Global image of ozone field for Fig. 13 for 21 June 2016 at 18:41 UTC

670 f13 Longitudinal or diurnal variation of ozone for the Northern Hemisphere every 5° from 0° to 70° for
671 21 June 2016 at 18:41 UTC. The grey bands are the individual data points in the band. The solid lines are
672 a Lowess(0.05) fit to the data points representing a solar time average from 0.6 to 0.7 hours depending
673 on latitude. The SZA is limited to ±70°. Longitude = 0 Corresponds to 18:41 local time and longitude = -
674 180 corresponds to 06:41 local time.

675 f14 Global image of ozone field for Figs. 15 and 16 for 17 April 2016 at 18:36 UTC.

676 f15 Southern Hemisphere: Solid lines are approximately 30 minute averages in solar time at 18:38 UTC
677 on 17 April 2016 for ozone variation between 0° and 55° S latitude in 0.25° latitude bands for 17 April
678 2016 at 17:36 UTC.

679 f16 Northern Hemisphere: Solid lines are approximately 30 minute averages in solar time at 18:38 UTC
680 on 17 April 2016 for ozone variation between 0° and 75° N latitude in 0.25° latitude bands for 17 April
681 2016 at 17:36 UTC.

682 f17 Erythemal irradiances calculated from Eq. 13 and from the EPIC ozone and LER data obtained on
683 April 17, 2016 at 18:35 UTC. The scale shows both the irradiance values in W/m² and the UV index
684 ranging from 0 to 20. This scene is centered over the Pacific Ocean and shows a peak UV index of about
685 15. Since this period is close to equinox, the sun is nearly overhead just north of the equator with solar
686 noon at 98.75° W longitude and overhead near 10° N.

687 f18 Erythemal irradiances centered over South America on November 23, 2015 at 16:19 UTC showing
688 extremely high values in the Andes Mountains in Peru, Bolivia, and Chile corresponding to a UV index
689 greater than 20. Local solar noon is at 64.75° W and overhead near 20° S.

690 f19 Erythemal Irradiances in a longitudinal slice at 20° S through a peak occurring in the Andes
691 mountains. Local noon is at 64.75° W.

692 f20 Erythemal irradiances centered over the United States on June 21, 2016 showing high values over
693 the Rocky Mountains and a portions of the Sierra Nevada Mountains. The UV index reaches about 15.
694 Local solar noon is at 99.75° W and overhead near 23.3° N.

695 f21 Erythemal UV irradiances centered over the Indian Ocean on June 21, 2016 showing high values over
696 the Himalayan Mountains with the UV index exceeding 14. UV levels are moderated by partial cloud
697 cover reflection of radiation back to space. Solar noon is at 80.25° E.

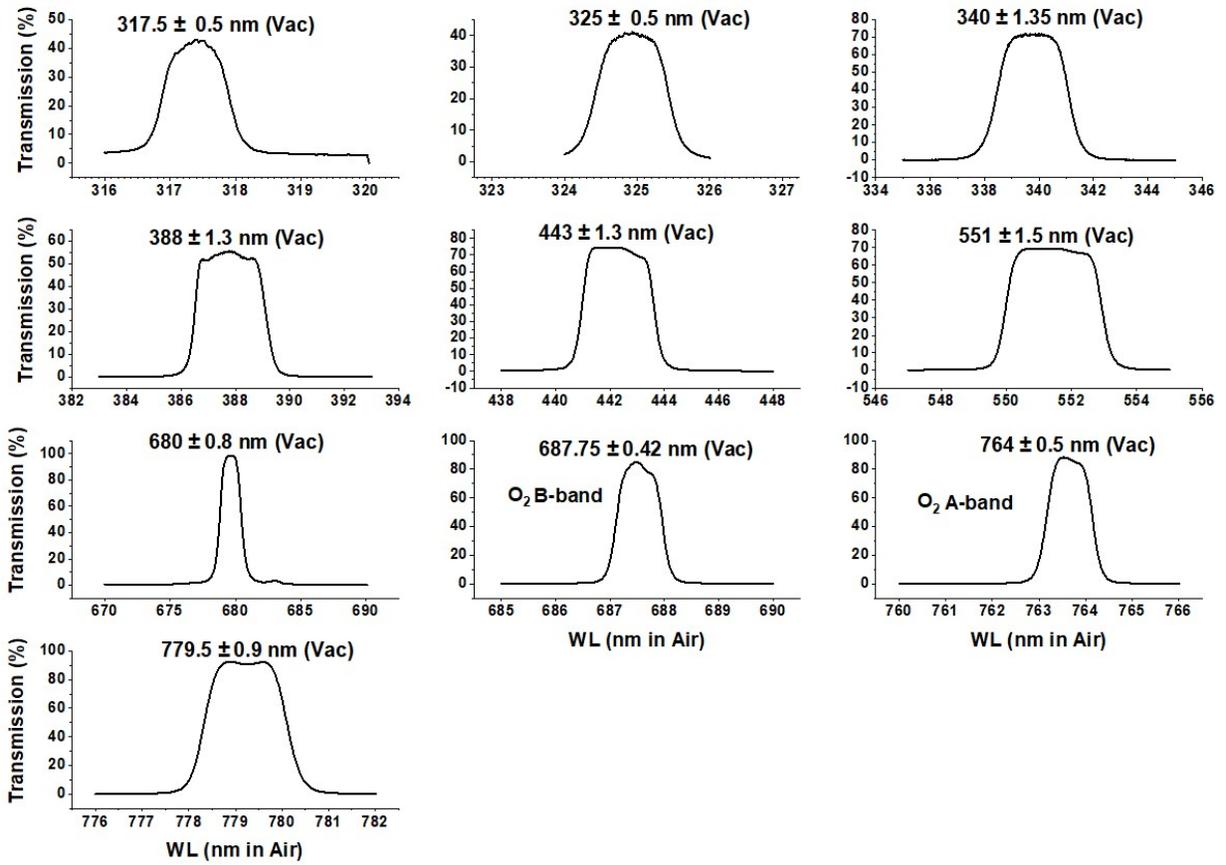
698 f22 Erythemal Irradiances in a longitudinal slice at 32° N through a portion of the Himalayan mountains.
699 Local solar noon is at 80.25° E.

700 fA1 An illustration of DSCOVR's Lagrange-1 orbit

701

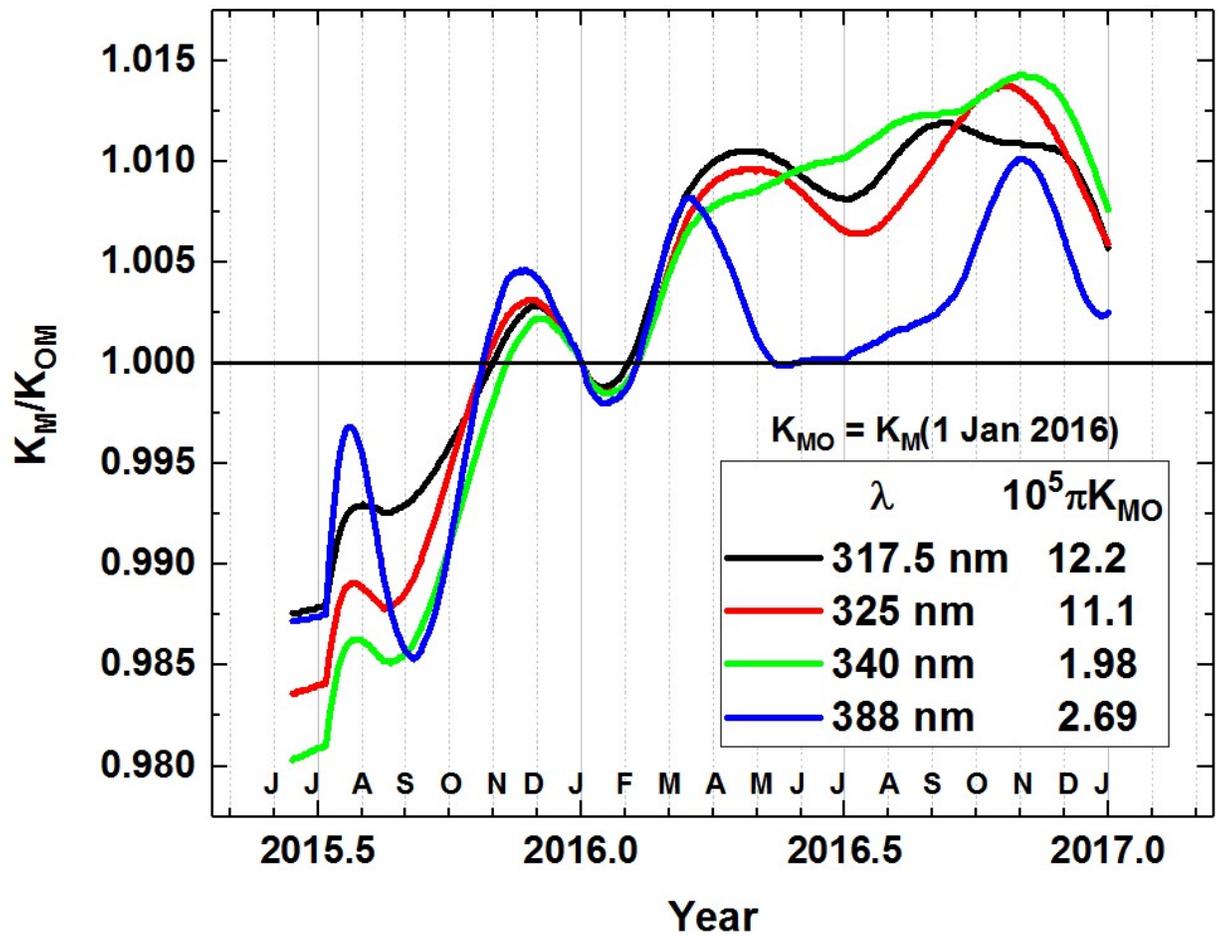
702 **Figures**

703



f01

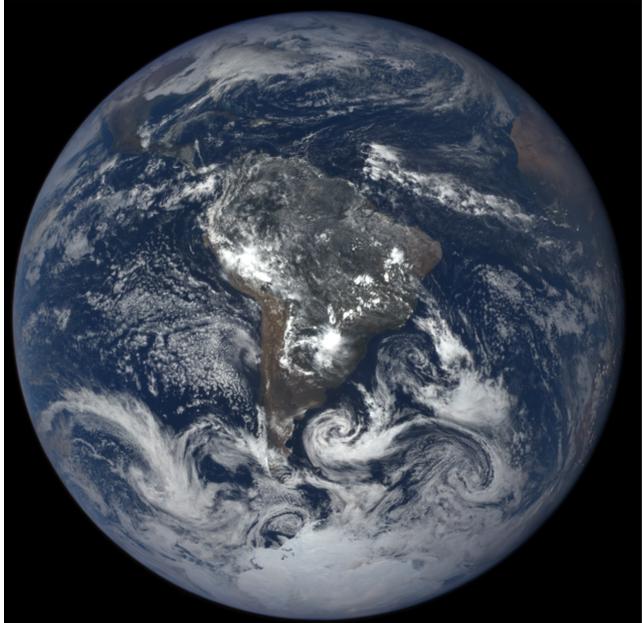
704



f02

705

706

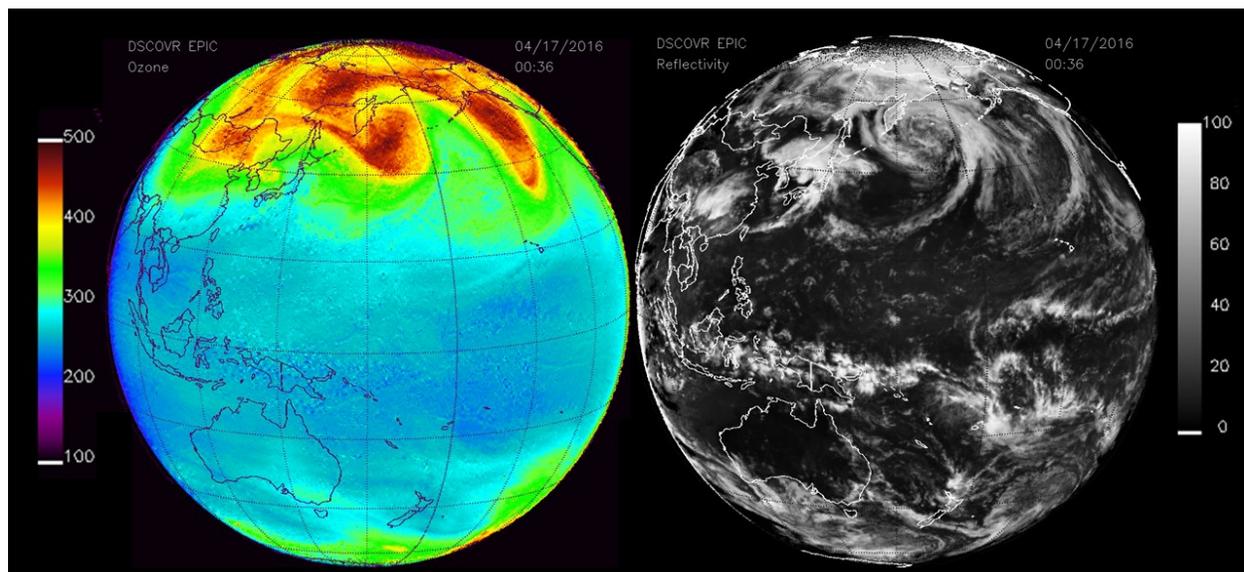


f03

707

708

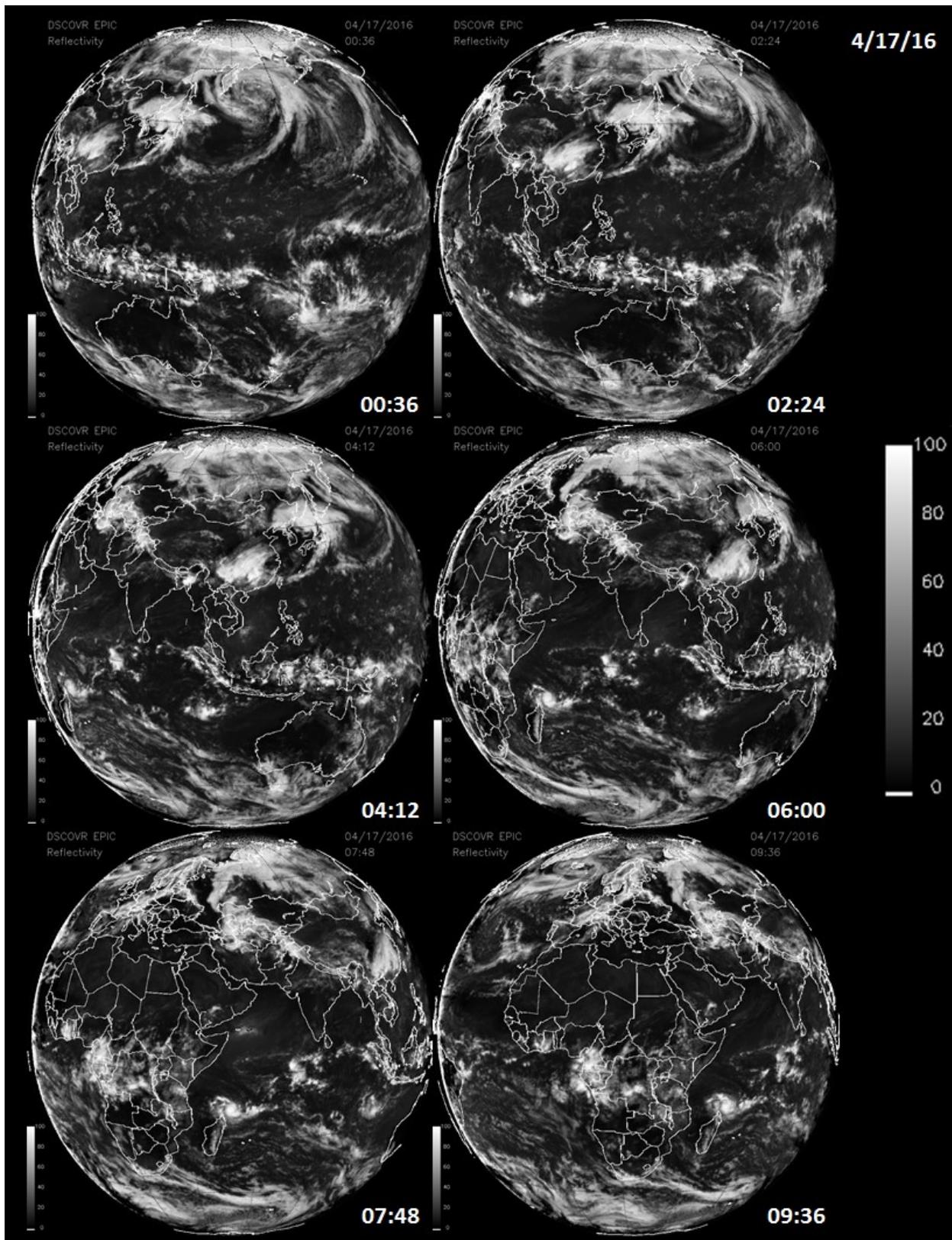
709



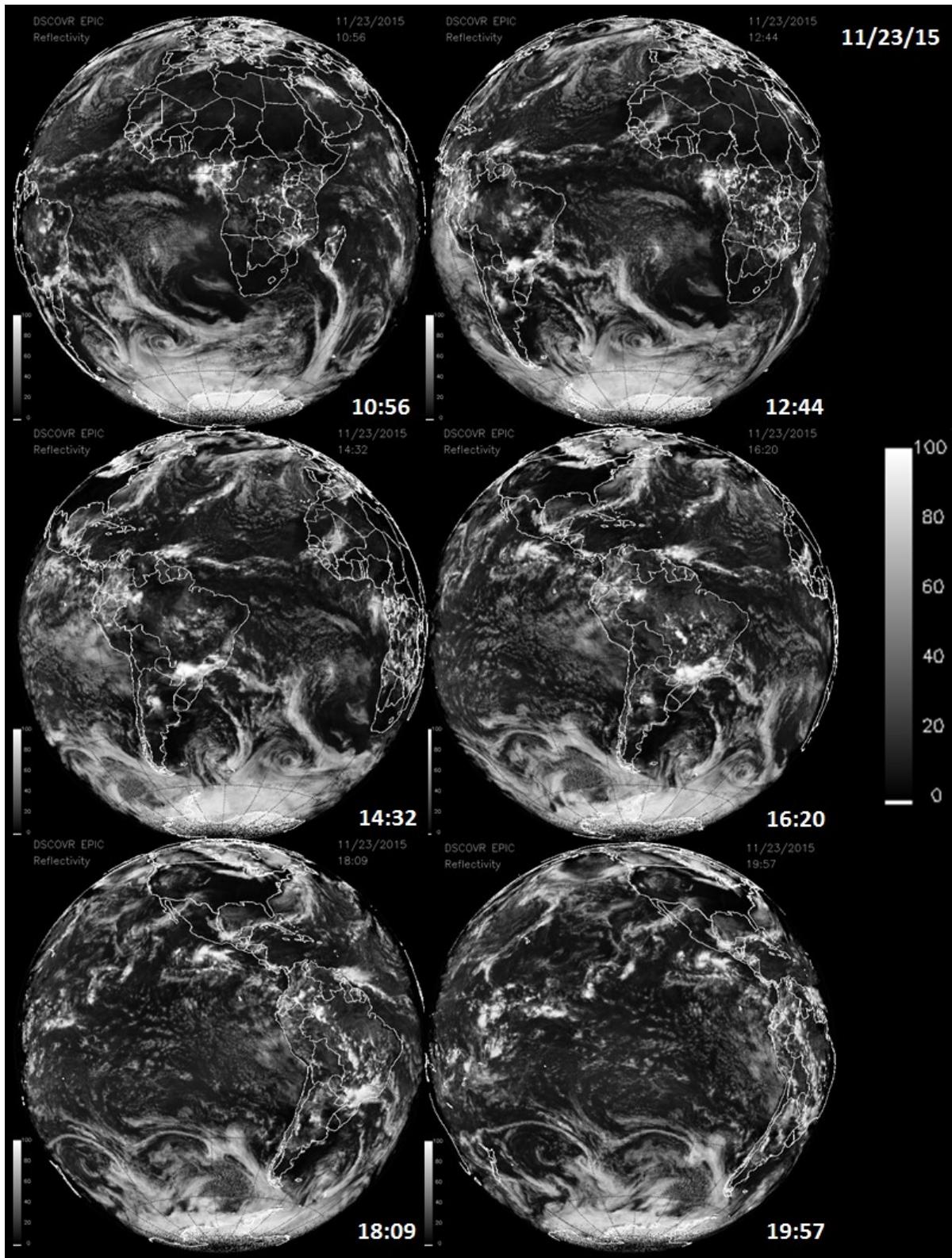
f04

710

711



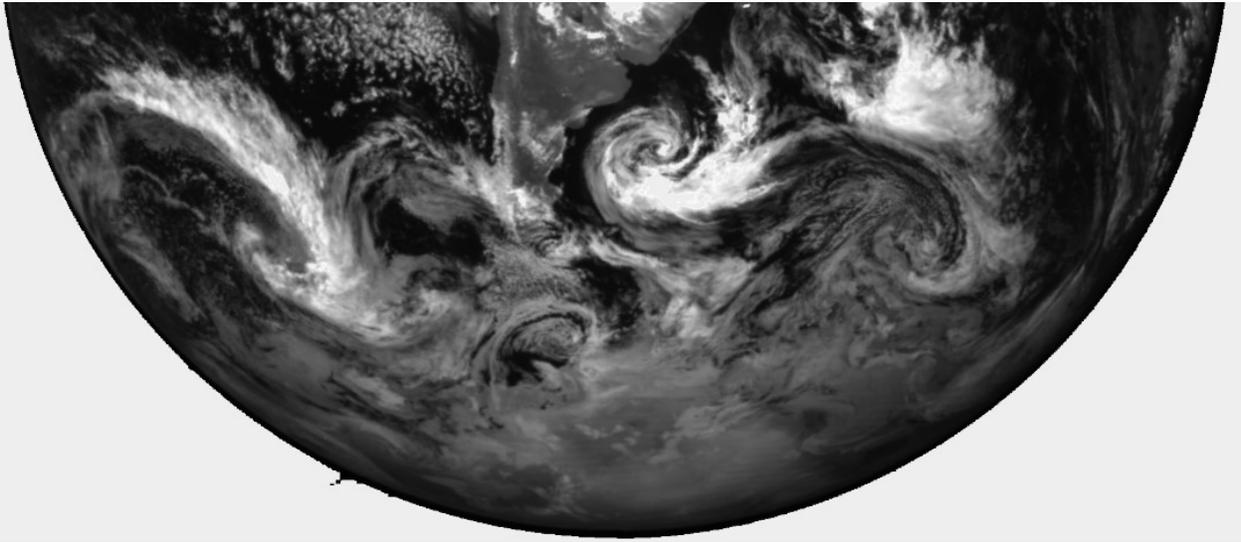
f05



f06

712

713



f07

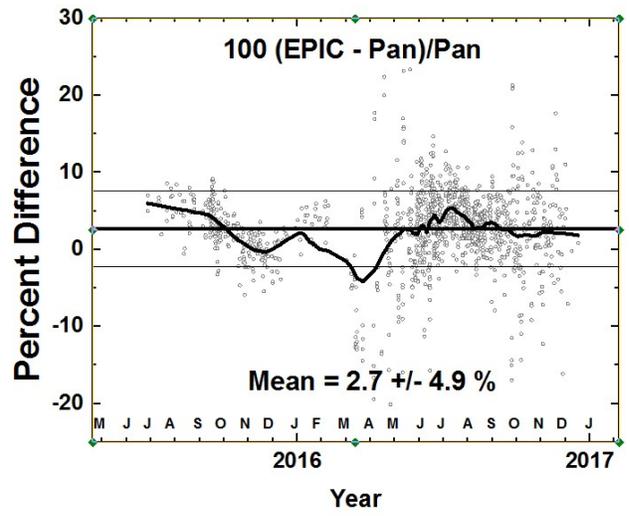
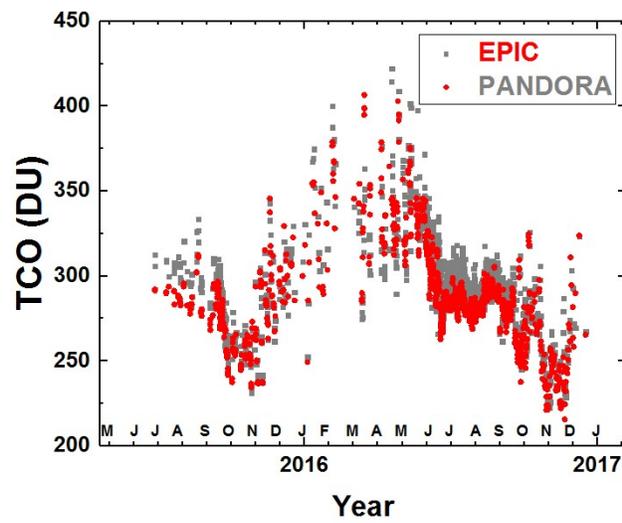
714

715

716

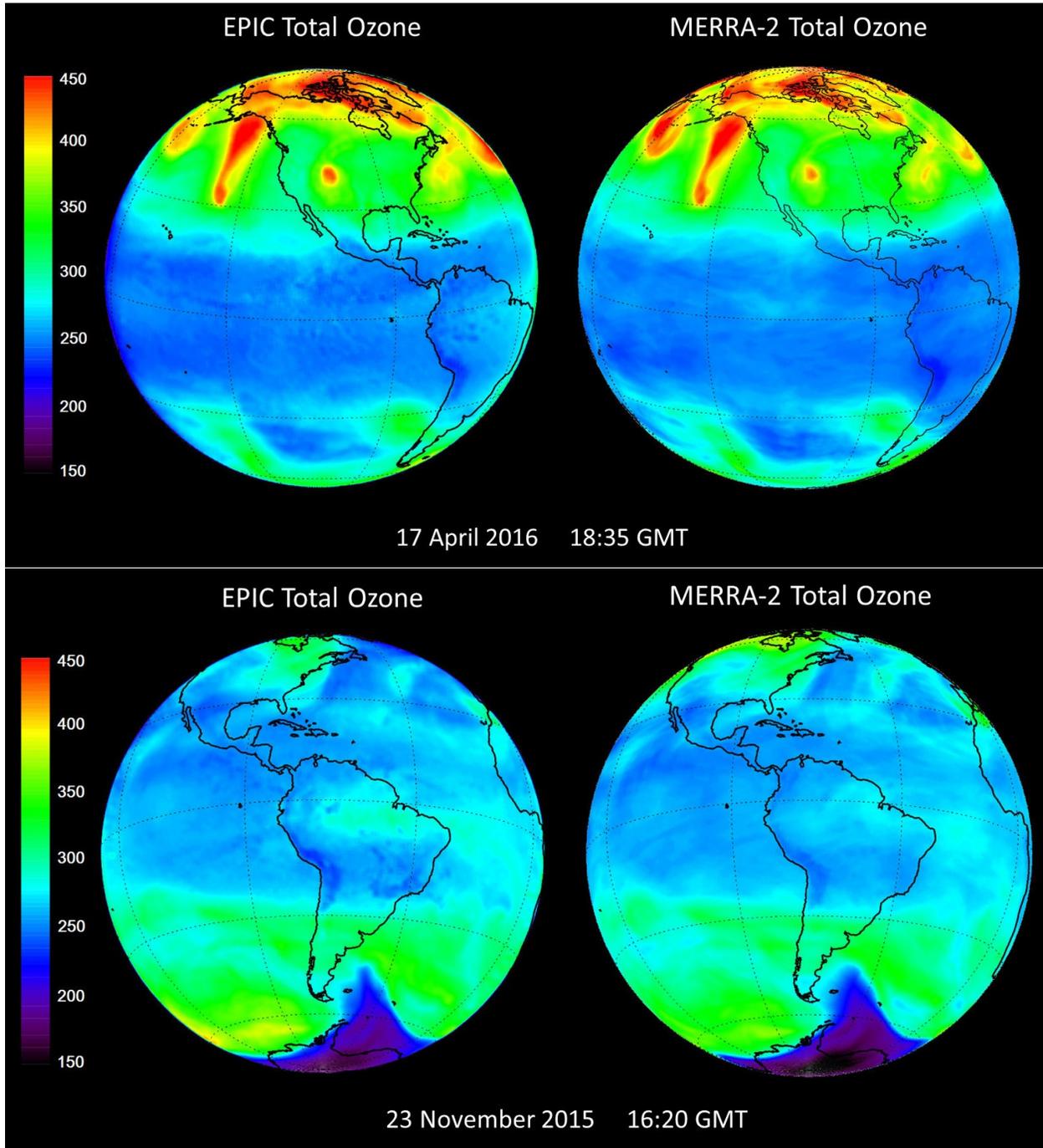
717

718

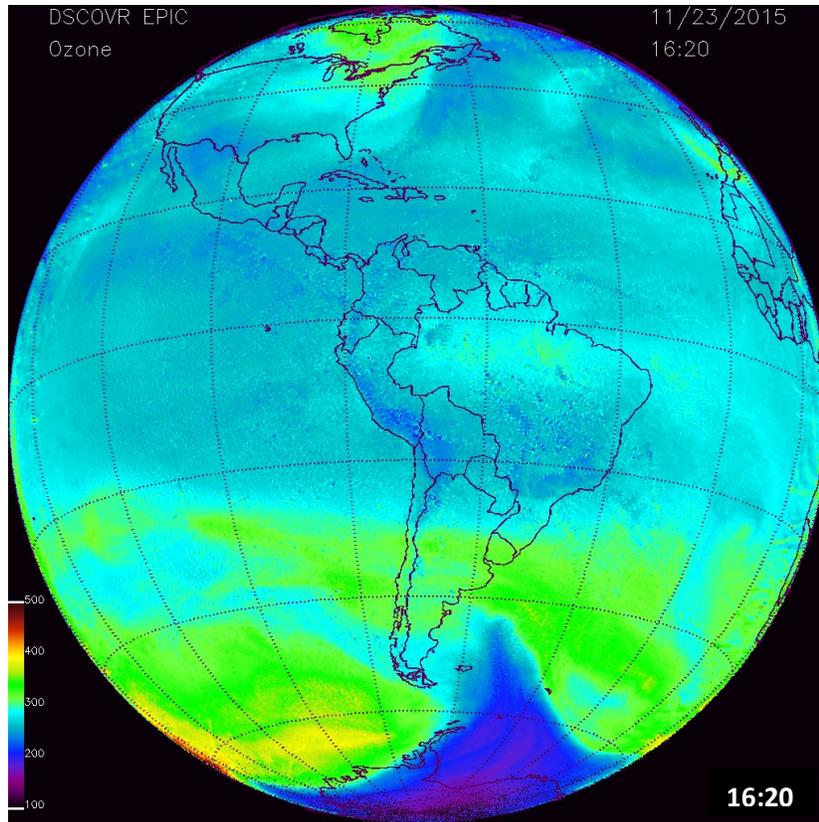


f08

719



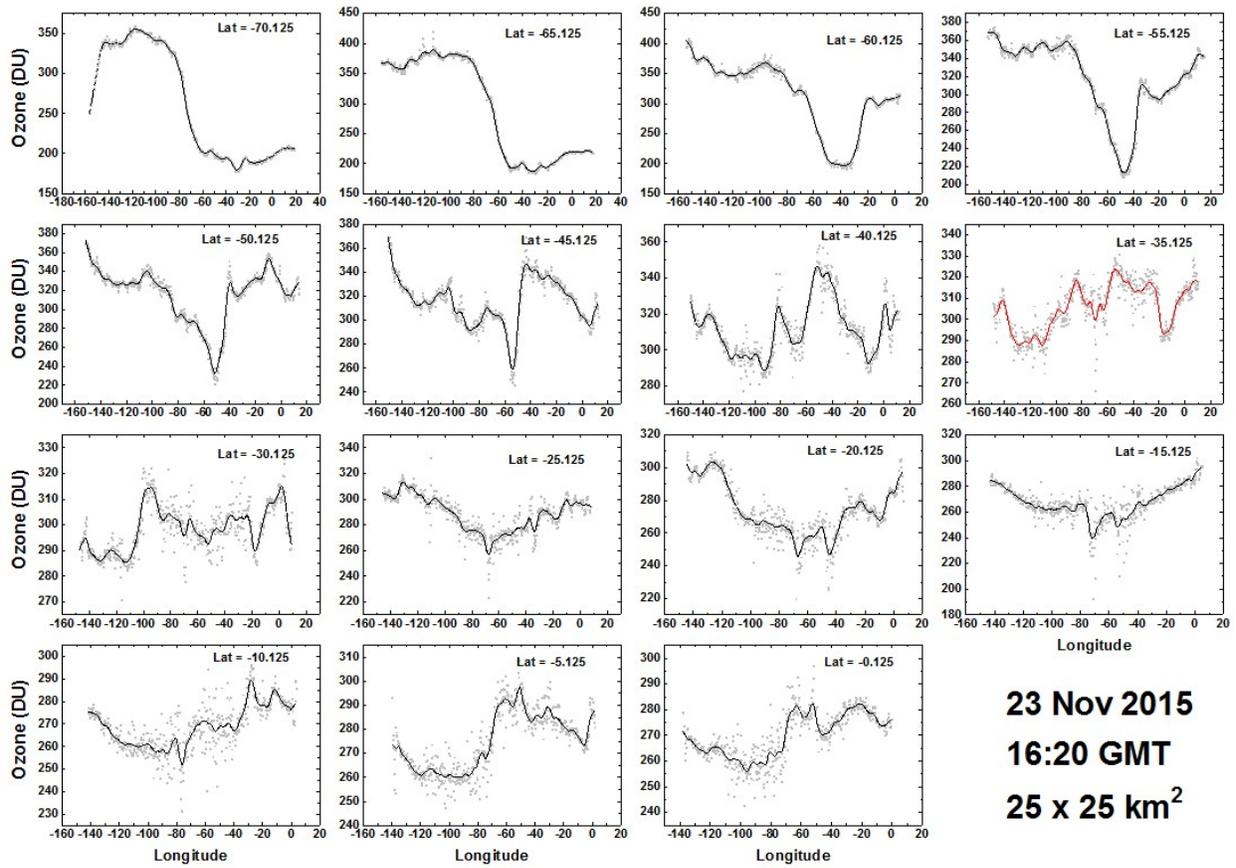
f09



f10

722

723

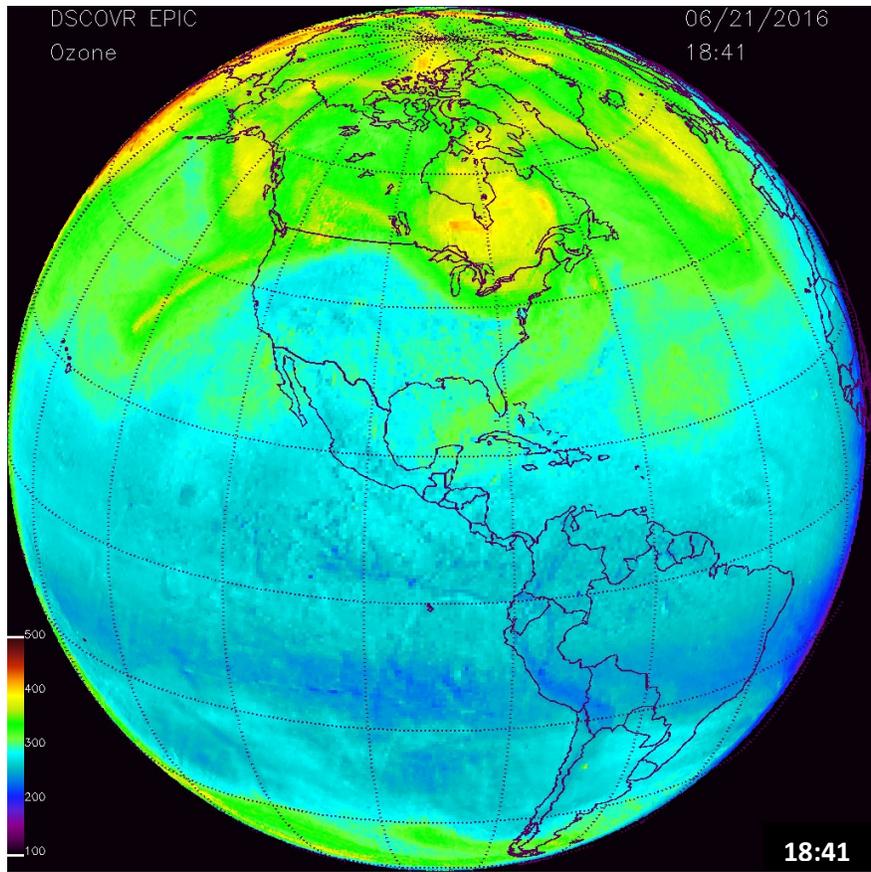


f11

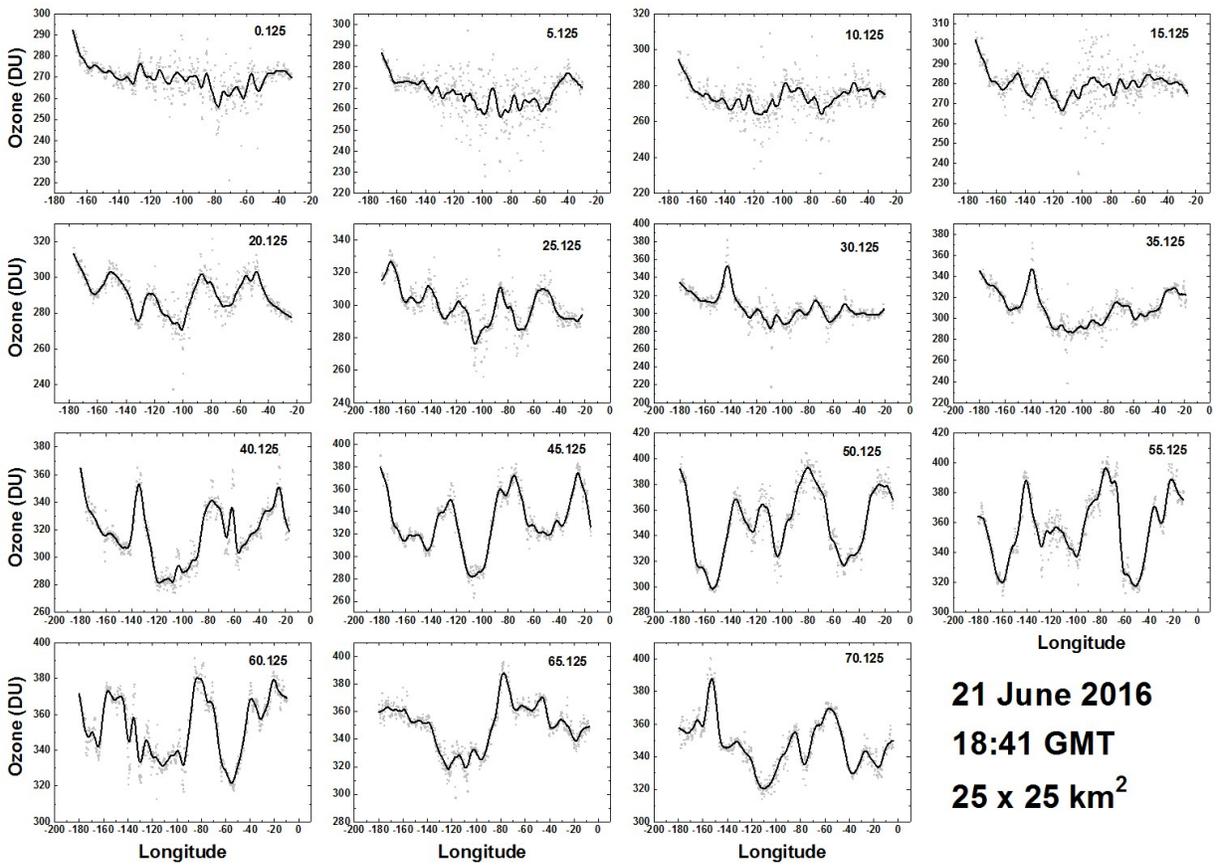
724

725

726
727
728
729
730
731
732
733
734
735
736
737
738
739
740
741



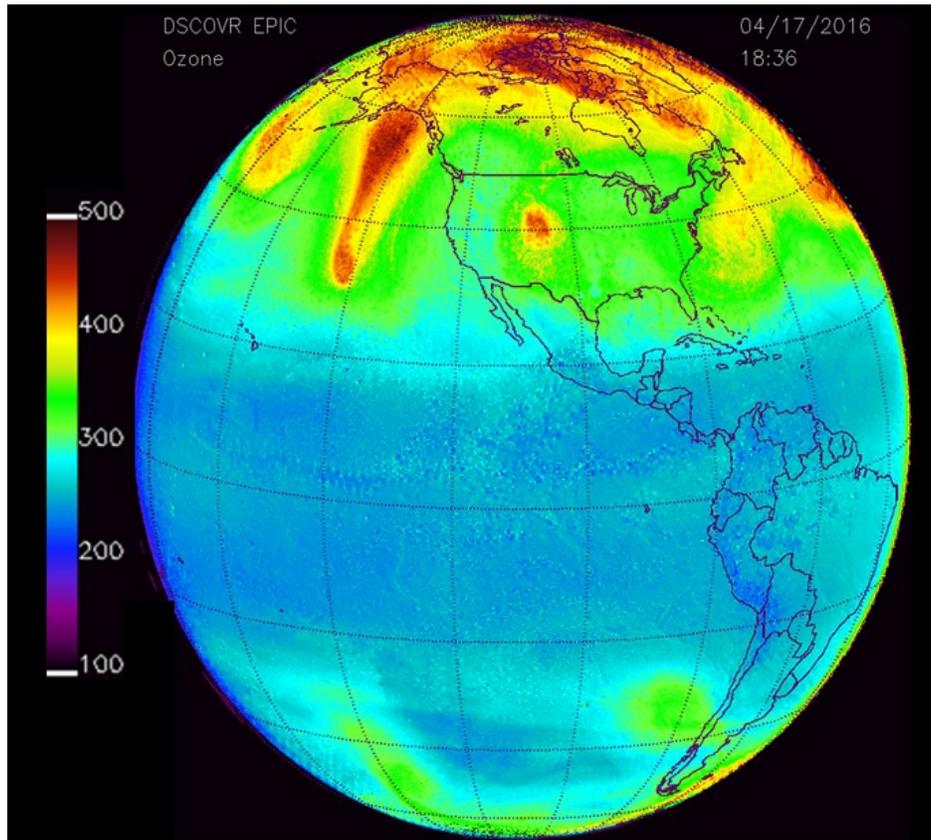
f12



f13

742

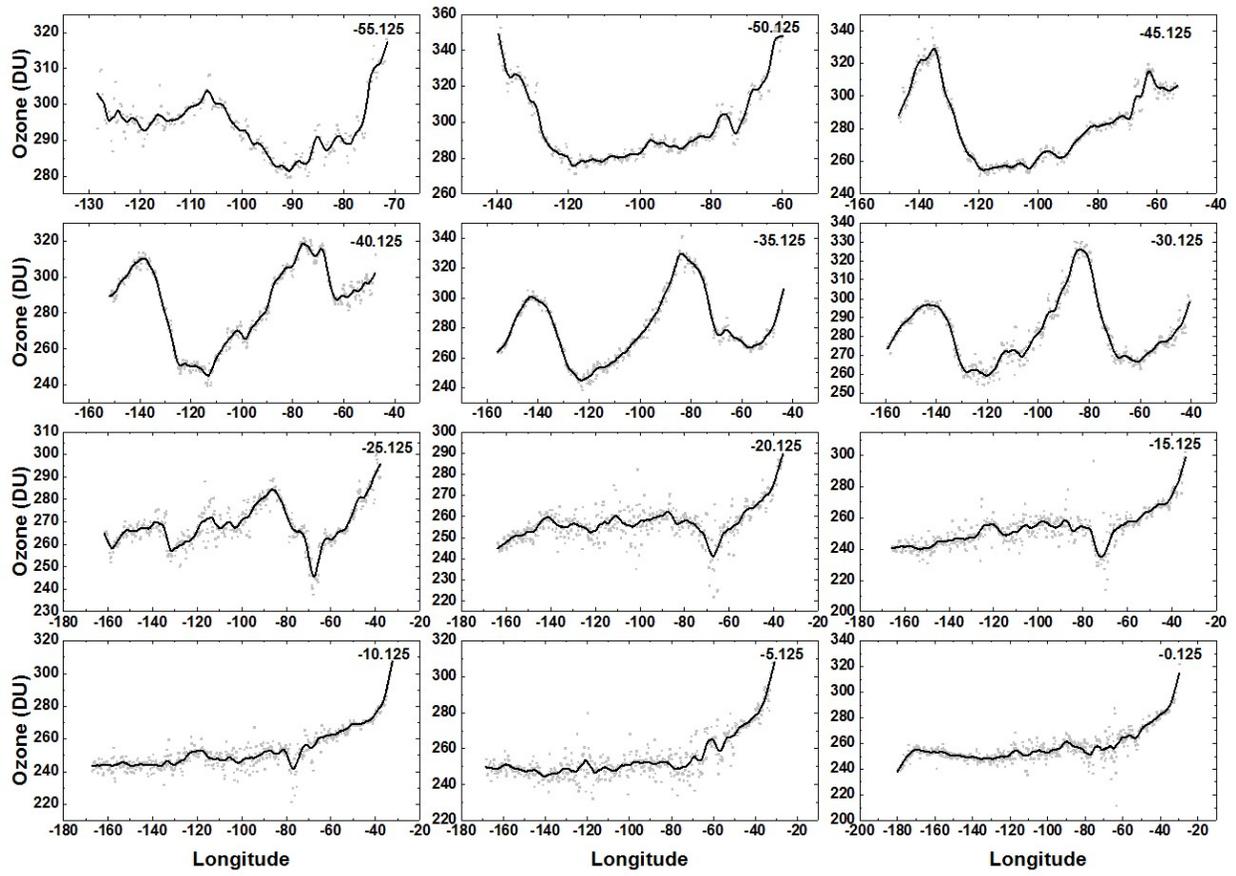
743
744
745
746
747
748
749
750
751
752



f14

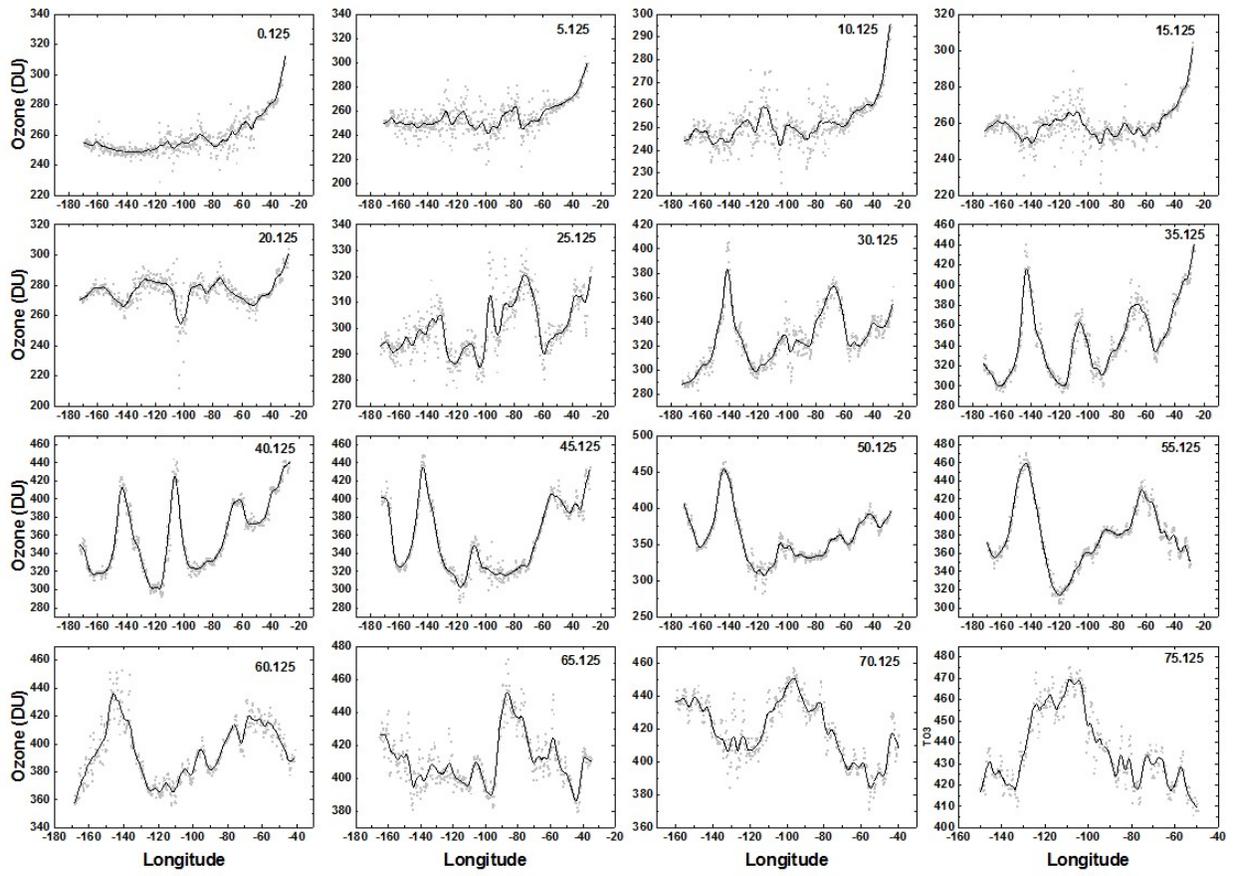
753

754



f15

755

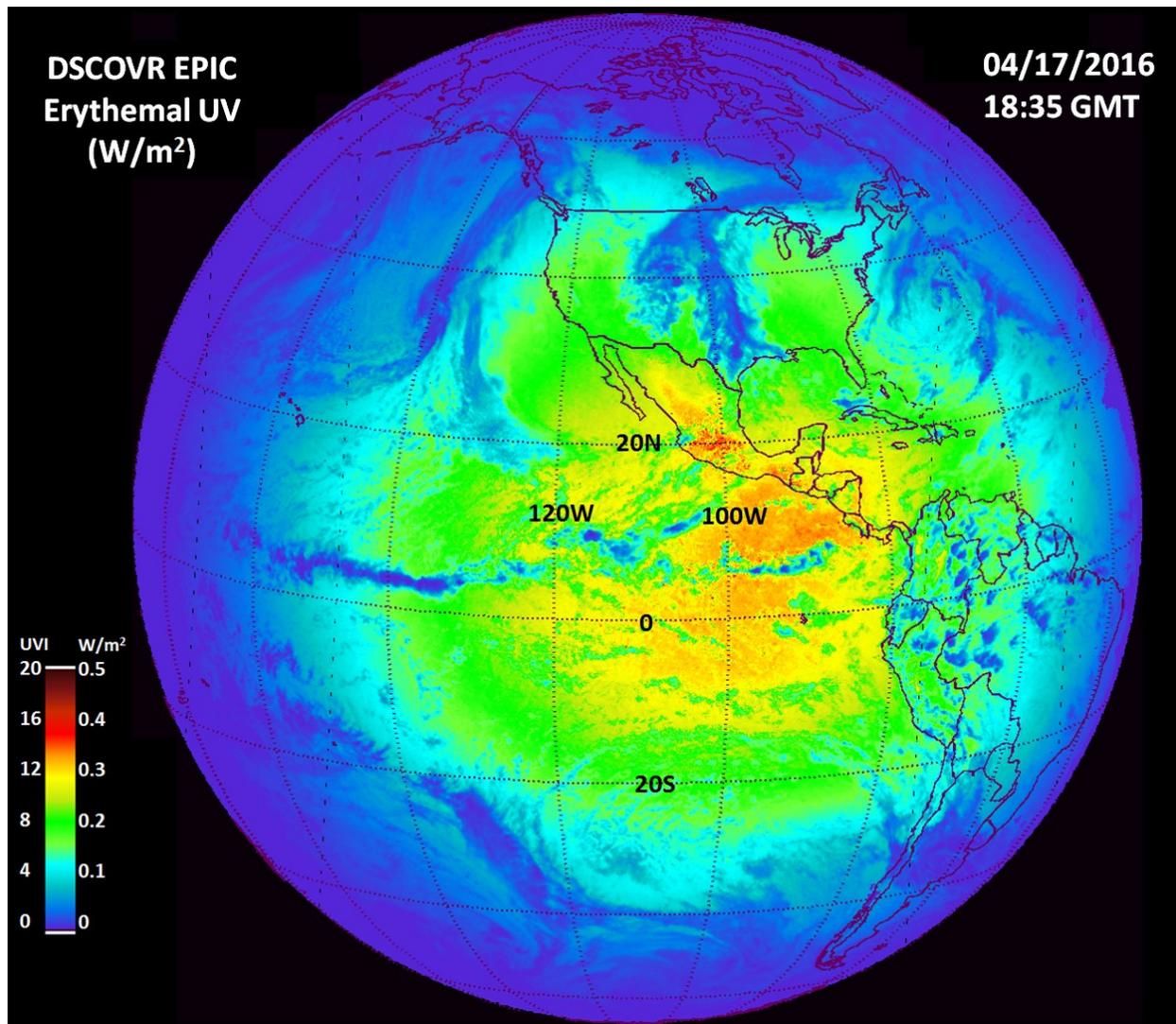


f16

756

757

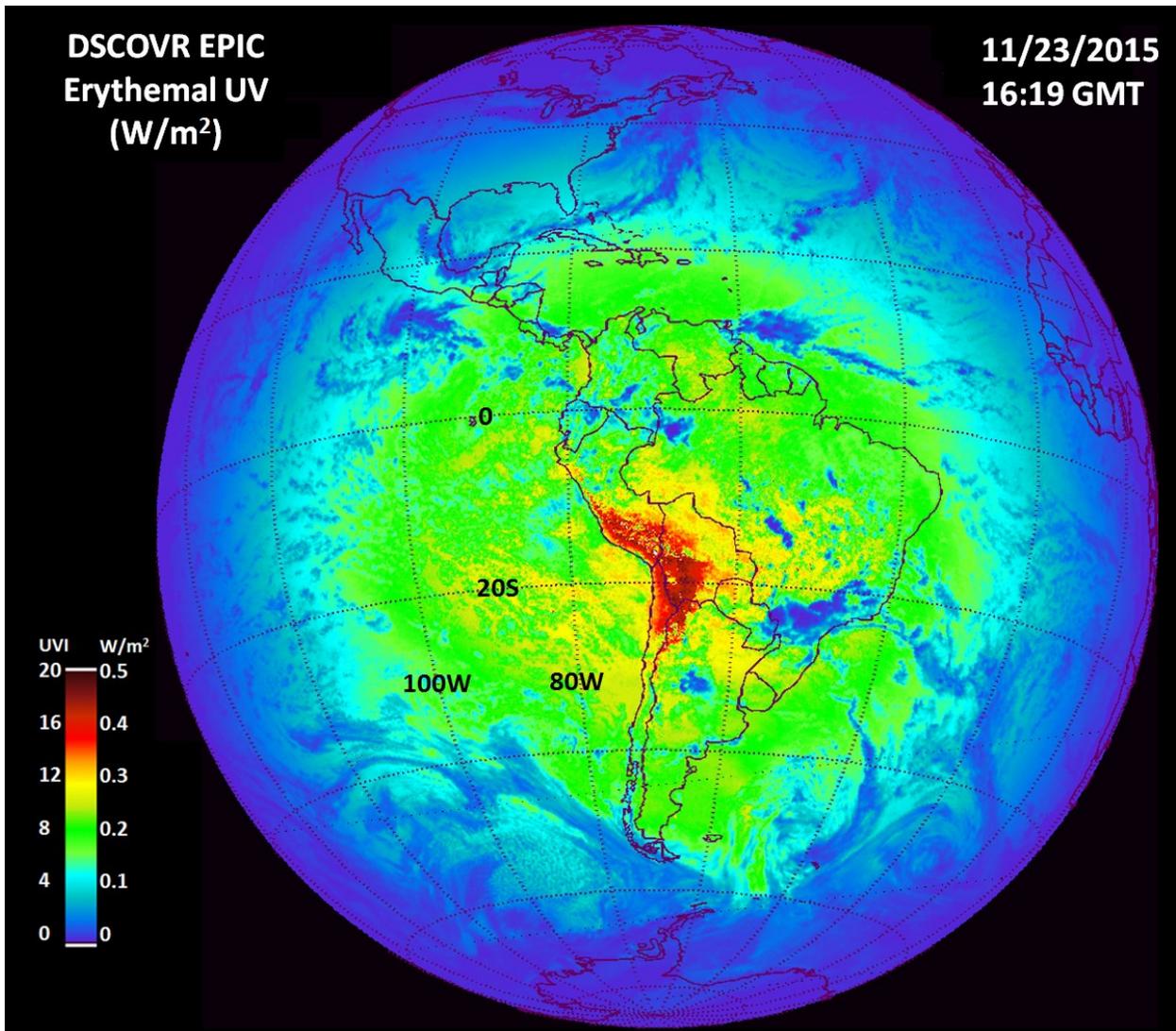
758



f17

759

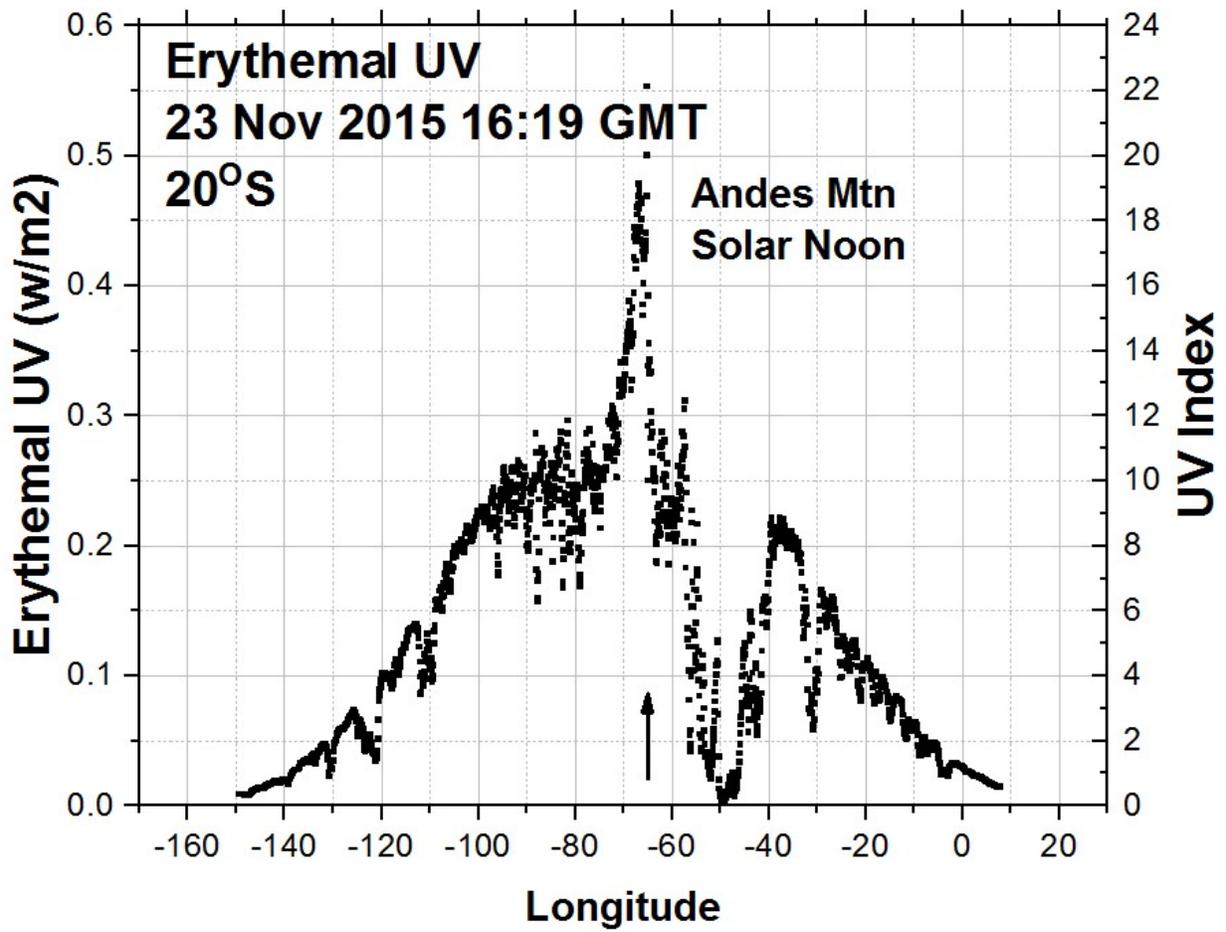
760



f18

761

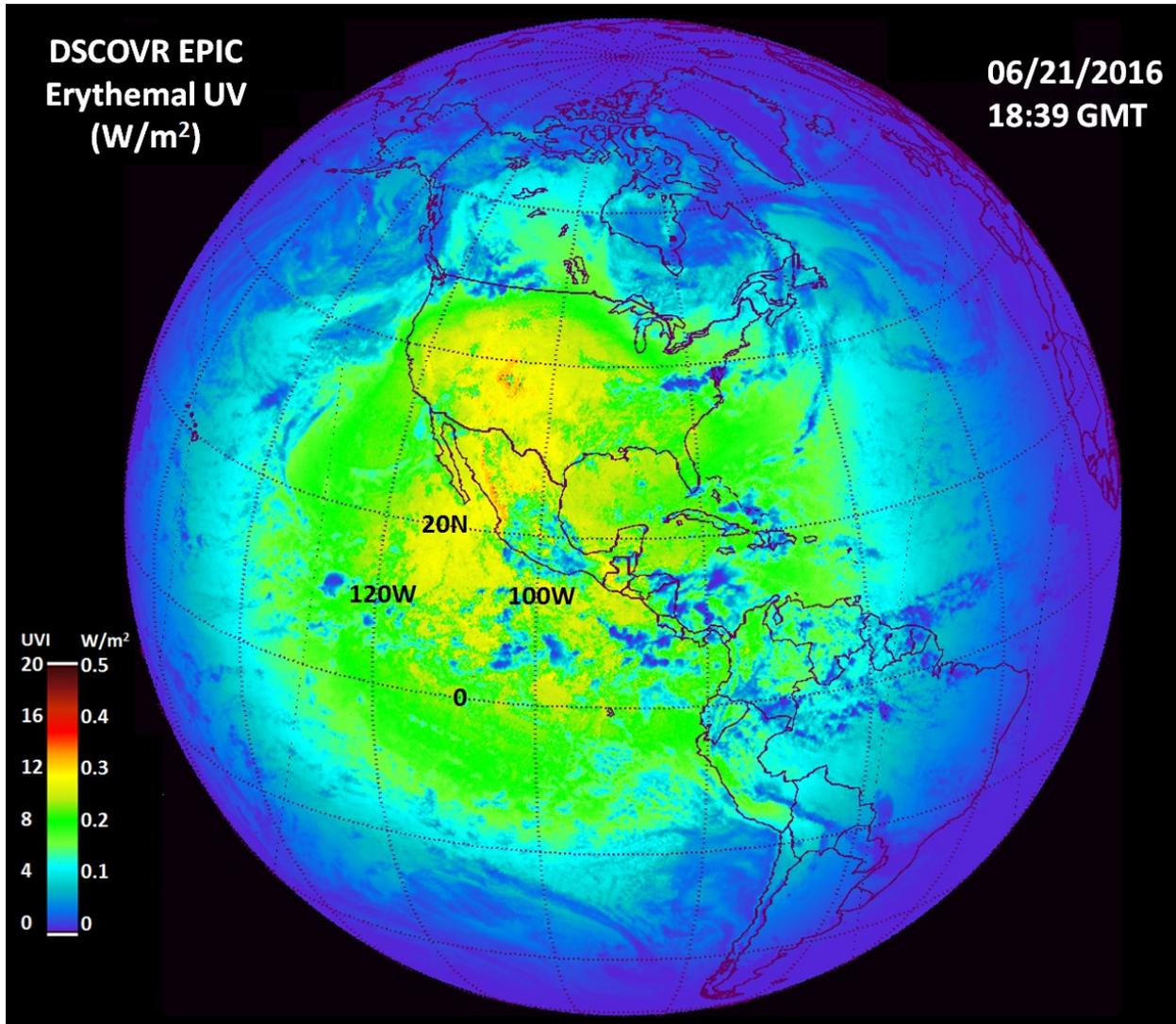
762



f19

764

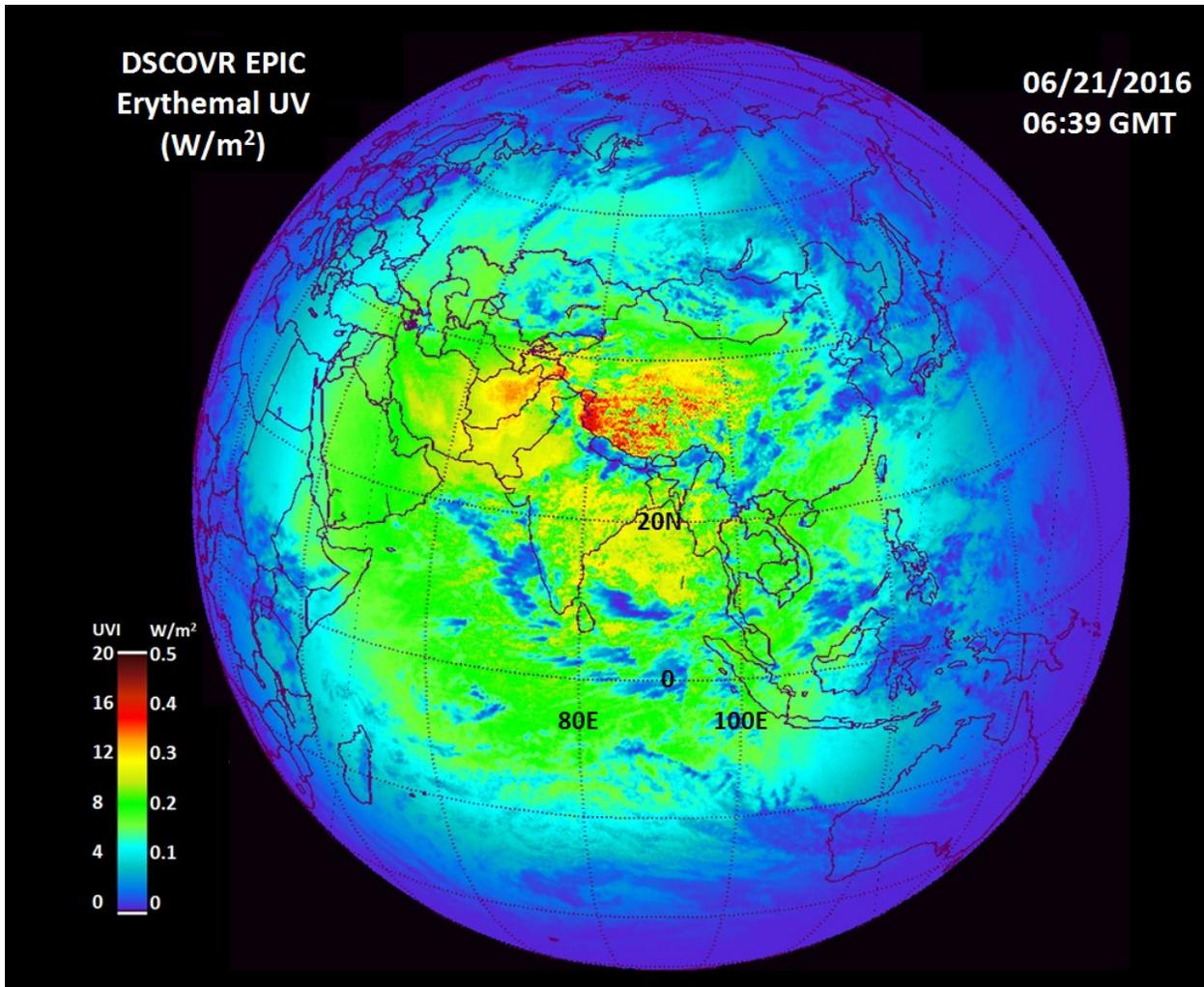
765



f20

766

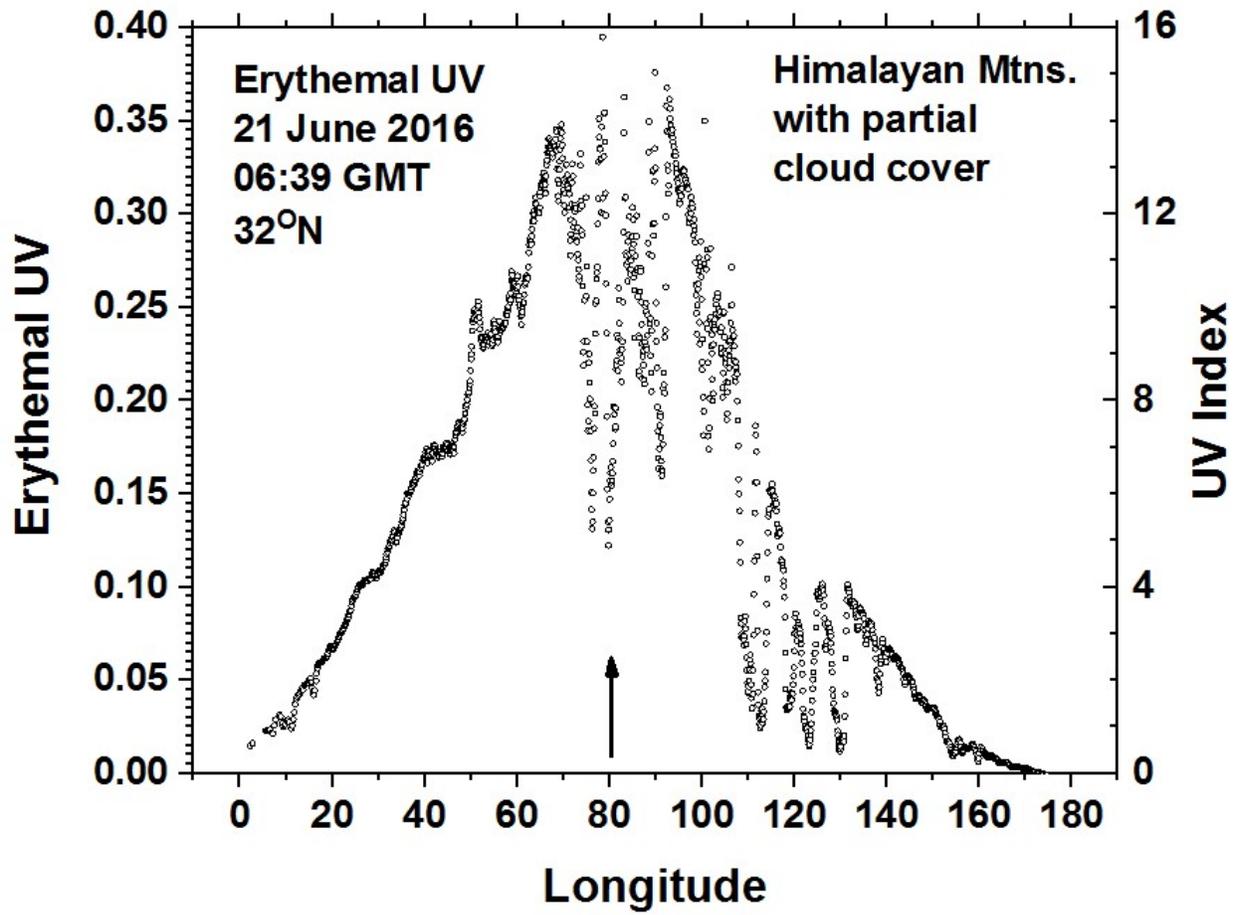
767



f21

768

769



f22

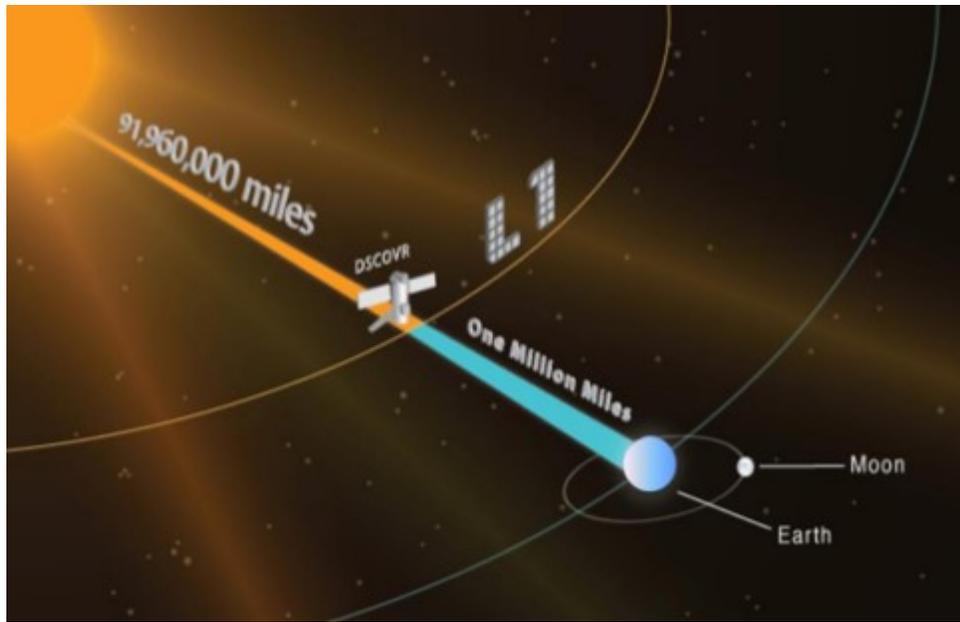
770

771

772

773

774



fA1

775

with CIE Erythemal.

p.15, line 510: ensure Careless. Now reads “ EPIC’s synoptic measurements ensure that”

p.24, line 641: their Now reads, “Normalized calibration functions referenced to their value at 4 Jan 2016”

p.24, line 644: 6 months Now reads, “p.24, line 644: “6 months apart”

Formatting does not fit yet AMT style. I need guidance here from the editor. I will read through the author instructions again.

Interactive comment on Atmos. Meas. Tech. Discuss., doi:10.5194/amt-2017-155, 2017.

Printer-friendly version

Discussion paper

