Journal of The Korean Astronomical Society
36: S83 ~ S91, 2003

SUNSHINE, EARTHSHINE AND CLIMATE CHANGE: II. SOLAR ORIGINS OF VARIATIONS IN THE EARTH'S ALBEDO

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ABSTRACT

There are terrestrial signatures of the solar activity cycle in ice core data (Ram & Stoltz 1999), but the variations in the sun's irradiance over the cycle seem too small to account for the signature (Lean 1997; Goode & Dziembowski 2003). Thus, one would expect that the signature must arise from an indirect effect(s) of solar activity. Such an indirect effect would be expected to manifest itself in the earth's reflectance. Further, the earth's climate depends directly on the albedo. Continuous observations of the earthshine have been carried out from Big Bear Solar Observatory since December 1998, with some more sporadic measurements made during the years 1994 and 1995. We have determined the annual albedos both from our observations and from simulations utilizing the Earth Radiation Budget Experiment (ERBE) scene model and various datasets for the cloud cover, as well as snow and ice cover. With these, we look for inter-annual and longer-term changes in the earth's total reflectance, or Bond albedo. We find that both our observations and simulations indicate that the albedo was significantly higher during 1994-1995 (activity minimum) than for the more recent period covering 1999-2001 (activity maximum). However, the sizes of the changes seem somewhat discrepant. Possible indirect solar influences on the earth's Bond albedo are discussed to emphasize that our earthshine data are already sufficiently precise to detect, if they occur, any meaningful changes in the earth's reflectance. Still greater precision will occur as we expand our single site observations to a global network.

Key words : Earth : albedo, global climate — Moon: reflection

I. INTRODUCTION

It is important to understand the origin of terrestrial signatures of the solar activity cycle, as seen in ice core data going back more than 100,000 years (Ram & Stoltz 1999). In all probability, such knowledge is key to determining whether there is an on-going global change in the earth's climate. To learn this, one needs precise, global/integrated measures of relevant quantities. The earth's climate is driven by the net sunlight deposited in the terrestrial atmosphere, and so, is critically sensitive to the solar irradiance and the earth's albedo. Variations in the sun's output over a solar cycle are too small to leave a signature (Lean 1997; Goode & Dziembowski 2003), so we need to carefully consider the earth's reflectance, which we have been measuring from Big Bear Solar Observatory for several years. The albedo shows considerable variation, Goode et al. (2001).

The power going into the earth's climate system is

$$P_{in} = C\pi R_e^2(1 - A), \tag{1}$$

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where C is the solar constant, R_e is the earth's radius and A is the short-wavelength (sunlight with a temperature of about 6000 K) Bond albedo (reflectance). Subsequently, this incoming power is re-radiated back into space at long-wavelengths (earthlight with a mean wavelength of about 15 μ or a temperature, T_e of about 255 K, the earth's temperature), where

$$P_{out} = 4\pi R_e^2\sigma\epsilon T_e^4 = 4\pi R_e^2\sigma\epsilon_a T_s^4, \tag{2}$$

where σ is the Stefan-Boltzmann constant and ϵ is the emissivity of the atmosphere (about 5.5 km high, where the long-wavelength radiation is emitted). Further, we have related the outgoing power to the global averaged surface temperature, T_s. For this, one must introduce an effective atmospheric emissivity, ϵ_a. Following Ramanathan et al. (1989), we define $g = 1 - \epsilon_a$, where g is the normalized greenhouse effect.

If our planet is in radiative equilibrium, $P_{in} = P_{out}$, and we have

$$T_s^4 = \frac{C}{4\sigma(1-g)}(1 - A). \tag{3}$$

This means that the Bond albedo directly controls the earth's mean surface temperature. Global warming would result if A and/or $1 - g$ decreased. By measuring the earth's reflectance and the spectrum of the reflected light, one can determine A and g, respectively.

Thus, it is fair to expect that global changes in the earth's climate would be manifest in changes in the earth's albedo. Here, we focus on a terrestrial determination of the earth's global albedo from an old, and largely forgotten method. That is, global albedo can be determined by measuring the amount of sunlight reflected from the earth and in turn, back to the earth from the dark portion of the face of the moon (the "earthshine" or "ashen light"). The most important historical program of earthshine measurements was carried out by Danjon (1928, 1954) from a number of sites in France. He used a "cat's-eye" photometer to produce a double image of the moon, allowing the visual comparison of the intensities of two well-defined patches of the lunar surface — one in sunlight and the other in the earthshine — at various lunar phases. Using the "cat's-eye" mechanism, he stopped-down the light from the sunlit portion to match the brightness of the ashen portion. This differential measurement removed many of the uncertainties associated with varying atmospheric absorption and the solar constant, allowing Danjon to achieve his estimated uncertainty of roughly 5%, ignoring his appreciable systematic error from an incorrect determination of the moon's reflectivity. Our measurements are about an order of magnitude more precise than his, in large part because we have better measurement technologies available. We have also eliminated his large systematic error by correctly measuring the scattering from the moon as a function of the phase of the moon. At a better than 2% precision on individual nights and 0.5% on our best nights, our terrestrial measurements of the earth's albedo have a precision comparable to that from satellites.

We have been steadily observing the earthshine from Big Bear since 1998 to determine the earth's reflectance and its variations. Here we present and interpret the results of our observational work and simulations of the observations of the earth's reflectance, and briefly discuss our observations of the spectrum of the earthshine from the 60" telescope on Mt. Palomar.

II. THE ALBEDO OF THE EARTH

In Qiu et al. (2003) and Goode et al. (2003), we have detailed our method of determining the earth's reflectance from photometric observations of the bright (moonshine) and dark (earthshine) parts of the lunar disk. Here we discuss the results of several years of observations of the earthshine. In addition, we have simulated the reflectance of the earth, treating separately the parts in the earthshine and all of the earth in the sunshine throughout the hours of the day. In the simulations, we use scene models of the earth from the Earth Radiation Budget Experiment (ERBE) and real-time satellite cloud cover data. We discuss and compare the observational results with those of the simulations. Further, the seasonal changes in the earth's reflectance are contrasted and compared for the observations and the simulations.

On any one night, we determine the reflectance of most of the sunlit earth for a particular phase of the moon (the effective albedo). We need to integrate over all phases of the moon, θ, to determine a global or Bond albedo for the earth,

$$A = \frac{2}{3}\int_{-\pi}^{\pi} d\theta A^*(\theta) f_L(\theta) \sin\theta, \qquad (4)$$

where A^* is the effective, or average albedo associated with a particular night, $f_L(\theta)$ is the moon's Lambert phase function and A is the Bond albedo.

For convenience, the measured earth's albedo is often expressed as the magnitude of the effective albedo, A^*, i.e.,

$$[A^*] = -2.5\log A^*. \qquad (5)$$

This standard astronomical definition implies that the larger $[A^*]$ is, the smaller the albedo. Note that a 1% change in A^* corresponds to a about $0^m.01$ change in $[A^*]$.

III. DAILY OBSERVATIONS AND SIMULATIONS

In Figure 1, we show evening and morning earthshine observations overlaid on corresponding model calculations covering the entire day (for details, see Goode et al. 2003). The two lower panels show the earthshine as a function of time (note that the time is plotted in reverse chronological order). The solid curve shows the variation of the calculated effective albedo A^* during the twenty-four hour period, and the solid boxes are the observed effective albedo. These results come from near to a quarter moon and are compared with the earthwide cloud cover from the same day in the top panel.

The top panel shows the cloud cover maps illustrating which parts of the earth contributes to the earthshine. We highlight (the large bright areas) those parts of the earth that are the source of the earthshine (i.e., are simultaneously in the sunshine and are visible from the moon at some time during the observations). The cloud cover is also shown as secondary grayscale, for instance, the east-west dark bands just north and south of the equator illustrate cloudless areas. We also indicate with boxes an intersection of the earth's surface with the bisectrix of the spatial angle between the surface intersection of the lines from the center of the earth to the moon and the sun or, in other words, the point of equal angles, where the angle of incidence is equal to the angle of reflection.

The observations are consistent with the simulations for 24 March 1999, which is one of the nights for which the agreement is quite good. Perhaps the most striking

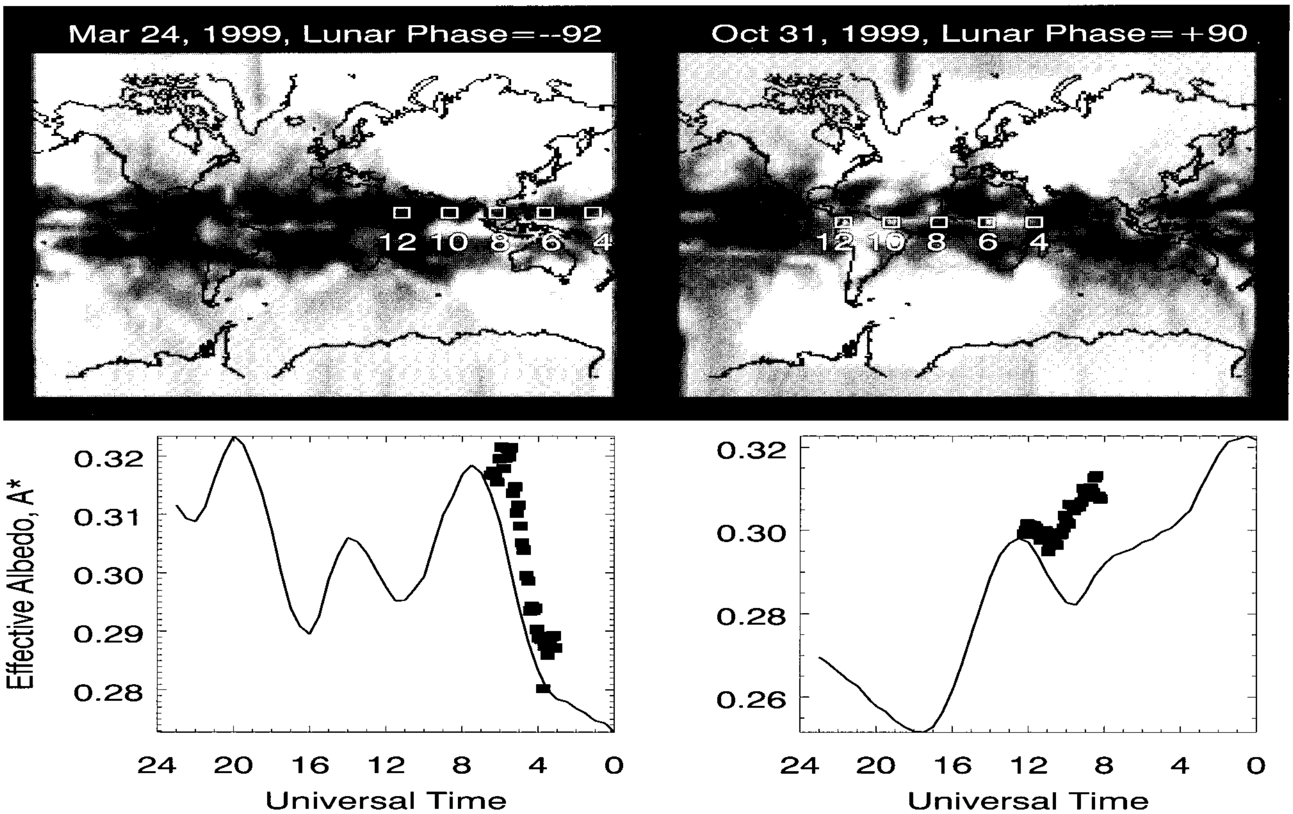

Fig. 1.— In the top panels, the extended bright areas highlight those parts of the earth that are the source of the earthshine. The satellite-derived cloud cover maps are shown in a secondary gray-scale with brighter areas indicating greater cloud cover. For 31 October 1999, note that the northernmost regions are not sunlit and the southernmost regions do not contribute to earthshine because the moon is fairly far north in the sky. The empty white boxes in the top panels indicate the longitudes of maximal contribution to the earthshine at the Universal Time (UT) shown. The solid boxes in the lower panels show the observed effective albedo as a function of time (note that the time axis is reversed), while the solid line indicates the simulated A^*.

aspect of the albedo for that date is that is it increases by more than 10% in about an hour, and this albedo is for the whole of the earth in the sunlight! On 31 October 1999, a more typical night, there is a discrepancy of about 5% relative (or 0.015 absolute) in the effective albedo. One may notice slight, and opposite, temporal offsets between the simulations and observations in the lower panels. One must bear in mind that the cloud cover data are a composite of many observations which are taken over about six hours (and sometimes up to 24). Thus, a precise timing between the observations and simulations is not possible. We simply assume that the cloud cover is invariant from one posting to the next and we make no effort to smooth the transition. If there were a rapid cloud formation or movement, it could result in observational and simulated results which do not have the same form. Still, the offset is a common feature. Typically, the effect will have one sign for an extended period of time and then it will change sign for another extended period. The origin of this behavior is discussed in the next section.

At high geographical latitudes above 45°, the cloud cover is fairly steady. Thus, the short timescale variations in reflectance in Figure 1 are due to primarily to irregularities in the fractional cloud cover at low latitudes and secondarily to the scene type. In the lower left panel of Figure 1, one sees that the observed and calculated maxima in A^* at 7:00 UT are due to a relative cloud excess over the Far East, while the calculated local minimum in A^* at 11:00 UT arises from the cloudless area above India and the Arabian Sea. In the lower left panel, the 10% change in A^* over about an hour is due to the increasing contribution of a cloudy Asia to the earth's reflectance as the earth rotates. An offset between the observed and calculated effective albedos is evident in the lower right panel. Typically, the observed results vary more about the mean than do the simulated ones. The larger variation in the observed results arises because the predicted seasonal variations are more muted.

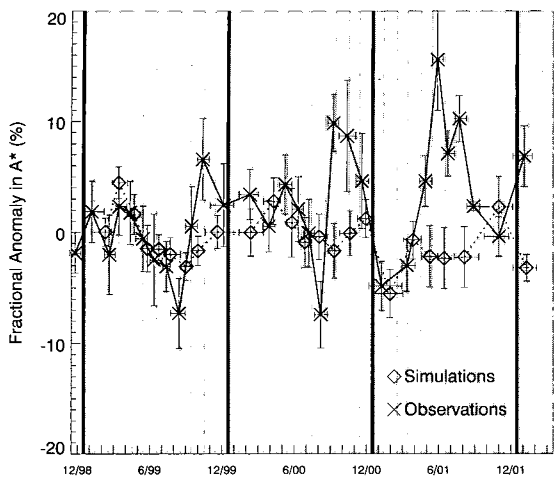

Fig. 2.— Seasonal anomalies in the effective albedo, A^*. From December 1998 through March 2002 , there are 340 nights of observations with 10 nights in each of the bins. From January 1999 onward, there are 268 nights for which we have both observations and contemporaneous satellite cloud cover data, which have been averaged in 24 bins with 10 nights in each. The $\times$'s show the mean of the observations, with the vertical bars being the standard deviation of the mean. The size of the latter stems from the large night-to-night variations in the cloud cover, rather than from uncertainties in the observations. The horizontal bars indicate the temporal span of each average. The diamonds indicate the corresponding simulation results. Anomalies are with respect to the mean for 1999.

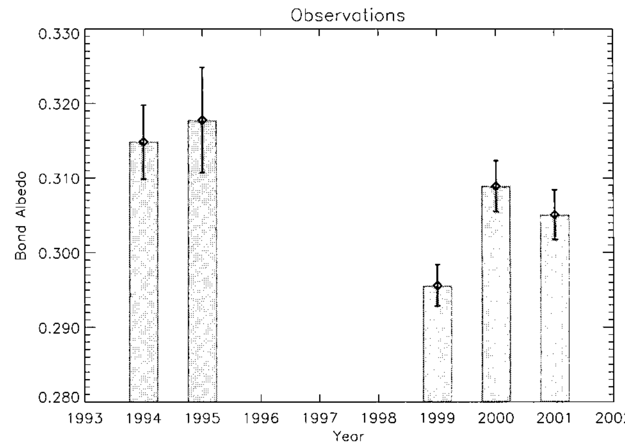

Fig. 3.— The earth's mean annual Bond albedo for each year in our observational earthshine record. The number of nights included in each year are tabulated in Table 1. Error bars represent $\pm 1\sigma$ deviation from the mean.

IV. SEASONAL AVERAGES OF THE EFFECTIVE ALBEDO

The seasonal variation of the earth's reflectance is not well-studied. In fact, Goode et al. (2001) have shown that over a year and a half (1999.0-2000.5) that the earth's seasonal variation was upward of 20%. This surprisingly large value is twice that determined from the simulations covering the same nights and the same parts of the earth. The formulation of the simulations was detailed in the previous section. Large scale, seasonal variations can be well-determined from the earthshine. The details of this determination are explained Goode et al. (2001, 2003). These seasonal anomalies, which are formulated as fractional changes in A^*, carry information about variations in weather, climate and surface type. The fractional seasonal variation of the earth's reflectance over three years (1999-2001), as determined from our earthshine observations, is shown in Figure 2.

One can see in Figure 2 a clear seasonal trend for 1999 and 2000, with the earth being brightest in the spring and fall generally, when it is also the cloudiest. We see the seasonal trends in spite of the fact that the night-to-night variations are a significant fraction of the seasonal trends. We emphasize that the large vertical error bars arise from the large night-to-night variations in the cloud cover, rather than any errors in the data. In fact, the night-to-night variations are large compared to the formal error bars for one night. With all of this, we see about 15-20% variation in A^* from season-to-season. However 2001 does not show the same seasonal pattern as 1999 and 2000. In the early months of the year, it shows a dip in the albedo that is also visible in the simulations, and it has a maximum in summer time, although this is not visible on the simulations.

The observations show about twice the variability as the simulations, with the differences being greatest at the extremes. The seasonal variation of albedo is the essential origin of the relatively greater variation among the observed $[A^*]$'s reported by Goode et al. (2001). They point out that muted seasonal amplitude of the simulations may well derive from the coarse binning of the scene models and/or the use of simulated snow and ice cover, but the treatment of the clouds is a stronger candidate. Our model contemplates only 13 different scenes and 4 cloudiness levels (0-5%,5-50%, 50-95% and 95-100%). Beyond the appreciable binning of the cloud cover, changes in cloud type will also affect the albedo, but it is not accounted for in the models.

V. INTER-ANNUAL CHANGES IN THE ALBEDO

In Qiu et al. (2003) and Goode et al. (2003), we developed and discussed a detailed methodology to measure and model the earth's Bond albedo. In Pallé et al. (2003), we focussed on probing the long-term changes in albedo and their global climatic impact. There we had more than 340 nights of observations covering the period from December 1998 to May 2002 (a time of high solar activity). We also have 73 nights of earthshine data covering 1994-1995 (a time of low solar activity).

For each of these nights, a mean effective albedo, A^*, is measured. A^* is the reflectance of the sunlit part of the earth visible from the moon at any particular lunar phase. However, to determine the Bond albedo of the earth, we need to integrate over all lunar phases (see Eq. 4). During the last three, year-to-year changes in the earth's albedo of order of 3-4% are observed. The results are plotted in Figure 3.

A series of straightforward corrections are applied to our daily [A*] measurements (see Qiu et al. 2003), so that our Bond albedo measurements are insensitive to natural variations, such as astronomical distances or lunar libration. Also, included is the systematic effect of the polar regions that are sometimes in the sunshine, but not in the earthshine. Since the polar regions are quite shiny, we must add 0.006 to the earthshine values. This point was developed in detail in Goode et al. (2003). Only the effect of lunar precession of the plane of the moon's orbit is not included, but this effect is quite small compared to, say, the effect of anisotropy (about five times smaller). The effect of precession serves to increase the measured Bond albedo from 1999 to 2000 and from 2000 to 2001. However, during this period, the effect is at least an order of magnitude smaller than the observed year-to-year changes (see Goode et al. 2003). Also, the effect cannot explain the change from 1994/95 to the present times, since at that earlier time the moon's orbital plane was approximately in the same precessional phase as during 1999/01.

The major change in albedo occurred between the early measurements and those that are the most recent. For the 1994/1995 period, we obtain a mean albedo of 0.310 $\pm$ 0.004, while for the more recent period, 1999/2001, the albedo is 0.295$\pm$0.002 (with a 0.6% precision in the determination). The combined difference in the mean A between the former and latter periods is -0.015$\pm$0.005, assuming the 1994/1995 and 1999/2001 uncertainties are independent. This corresponds to a 4.97% $\pm$ 1.66% decrease in the albedo between the two periods. We note that this result implies a positive climate forcing. One model that would be consistent with this result is that of Svensmark & Friis-Christensen (1997). It has been know for 50 years that the galactic cosmic ray flux is modulated by the solar cycle with the cosmic ray flux being maximal at solar activity minimum when the sun's shielding field is weakest. Svensmark and Friis-Christensen (1997) postulate that cosmic rays seed clouds and the result is a greater cloud cover, and hence albedo at minimum. They found a 3-4% greater cloud cover at minimum for half a cycle. Their picture is interesting but controversial. But is our result significant or correlated with the flux of galactic cosmic rays, the Zurich sunspot number (R_z) or the global mean temperature of the earth? Although we have only five annualized temporal points, a linear correlation analysis between the observed Bond albedo gives correlation coefficients of 0.34, -0.7 and -0.2, respectively. Thus, none reaches a 95% confidence level.

At this point, however, we stress that we do not have the same confidence in the mid-90's measurements, as we do in the more recent ones. So far, we have not been able to identify a reason, but we worry that there might be a calibration problem between the old and new data. In particular, a different field stop and a different camera was used in the 1994/95 telescope setup. In Qiu et al. (2003), the crucial importance of an accurate filter transmission determination was shown. Uncertainties in this area can produce strong offsets between different observations. For the 1994/95 data, we do not have enough nights to perform an accurate filter test, so we rely on the nominal transmission of the filters.

Could the difference between the two periods be terrestrial in origin? The years 1994/95 were in the midst of an El Niño event, while during the years 1999/01 a La Niña event was in progress (www.cdc.noaa.gov). The ISCCP global cloud coverage cloud indices seem to have an El Niño component with more clouds during the event. In principle, a greater cloud cover would imply a higher albedo, however there are strong regional trends (as opposed to global) in the cloud indices (Farrar 2000). A fact that argues against El Niño events being responsible for our higher albedo during the period 1994/95 is that our albedo simulations using ISCCP total cloud cover do not show a higher albedo during the period around 1998, when the strongest El Niño event on record took place (next section). In fact, the ISCCP-derived albedo for 1998 is lower than for 1994 or 1995.

So, we cannot confidently conclude whether the 5%$\pm$2% excess in the Bond albedo is a change tied to the systematic variation in solar activity, a spike in our record due to a relatively short-lived climate phenomenon like El Niño, a calibration problem, or most probably some combination of the three.

The uncertainty in the albedo determination strongly depends on the number of nights for which we have data. In this sense, we expect that increasing the number of earthshine stations (a process already underway), and judiciously spacing them in longitude will significantly reduce this uncertainty. The fact that the morning and evening earthshine measurements are almost identical (Goode et al. 2003) suggests that data from different stations may be easily combined to derive a 24-hour, annual Bond albedo with a precision far better than a percent, and monthly albedos with a precision closer to a percent. Further, we anticipate determining regional albedos, of large slices of the the earth, to about a percent. These will allow us to confidently distinguish between changes in the albedo arising from variations in solar activity from variations in weather phenomena, while reducing, or even eliminating calibration problems.

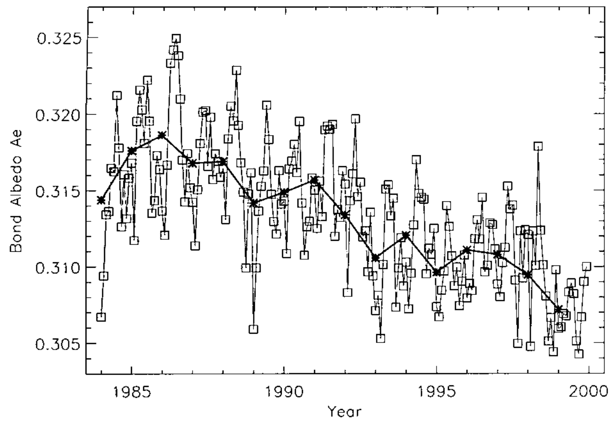

Fig. 4.— Using monthly average cloud cover data from the ISCCP data as input to the scene model, we calculate monthly averages of A that are represented by the boxes. We also show the annual averages of A as stars. The averages are plotted against time.

VI. SIMULATING THE EARTH'S ALBEDO, 1984-1999

In the previous section, we compared the observed effective albedo between the present time (1999-2001) of high solar activity and the most recent previous time of low solar activity (1994-1995). Here, we use the monthly averaged ISCCP-D2 data as input to compare and contrast our simulations of the albedo for the present solar cycle to those for the previous cycle. The simulations make use of cloud and snow and ice cover together with ERBE scene models to calculate a 30-minute resolution, global Bond albedo for the earth. In Figure 4, the earth's Bond albedo for the full period 1984-1999 is plotted. A general decreasing trend is apparent over the whole period. This decrease supports the observational result of a higher albedo in the period 1994/95 than there is at present. However, a solar cycle modulation of the albedo is not present. In particular, 1986 was a time of activity minimum, showing higher albedo than the 1990-91 maximum, but there is no increase in albedo going to the next activity minimum.

The lack of a solar modulation in the ISCCP total cloud cover data however, does not necessarily mean that there is no relationship between solar activity and cloudiness. There are many complex, and some apparently contradictory issues surrounding cloud cover detection. ISCCP D2 *total* cloud data uses visible (VIS) and infrared (IR) radiances to determine a percentage global coverage. Pallé & Butler (2000) already demonstrated that this parameter is correlated to solar activity only during the restricted period 1983-1991. However, ISCCP D2 products also include a distinction between high-, mid- and low-level cloudiness based on IR radiances only. When using these data, the correlation between low cloudiness and solar variability (or galactic cosmic rays) is maintained at least over the period 1983-1994. At the same time, high and mid-level cloudinesses are almost invariant during this period (Pallé & Butler 2000), so that the sum of the 3 IR cloud types does not equal the total cloudiness using IR+VIS radiances. ISCCP data are a combination of many satellite datasets subject to spurious changes and calibration problems. It is difficult then, not only to ascertain the reasons for the discrepancies, but also to decide which dataset should be used to study long-term trends in the earth's global radiation budget.

In the future, we are planning to include the distinction between cloud heights in our model parametrization. Low clouds are optically thick, and will have a bigger impact on the earth's albedo than mid- or high-level clouds. At present, we use only total cloud cover in our simulations (see Goode et al. 2003 for details of the models). It is quite possible then, that with the introduction of independent treatments of the different cloud types in our models, the Bond albedo variability will become more correlated with solar variability. Bluntly, there is an inconsistency between the treatment of the cloud cover that could, when resolved, yield model results having quite different temporal variation than appearing in Figure 4. The earthshine project circumvents all of these problems by directly measuring the earth's albedo without the need for parameterizing cloud quantities.

By measuring the albedo with our calibrated network, we will determine the true variation of the earth's albedo, and truly address questions like the variation of the albedo with changing solar activity.

VII. SOLAR CYCLE VARIATIONS IN THE EFFECTIVE ALBEDO?

There are many terrestrial signatures with an 11/22 year periodicity that, by default, would seem to be associated with the sun's magnetic activity cycle. Perhaps the most impressive is the demonstration of a wandering, near 11-year periodicity in the dust in Greenland ice core data going back more than 100,000 years (Ram & Stoltz 1999). With such signatures in mind, it is one of our primary goals to determine whether or not the earth's reflectance varies with solar activity. We have presented sufficient data to do a preliminary probe of this question. From observations and models, we conclude that the albedo has decreased from the period 1994/95 to 1999/01, however we have no compelling reason to attribute this decrease to changes in the sun.

Nonetheless, suppose the difference in reflectance were tied to the cycle. What size change in the net sunlight reaching the climate system would be caused by that change in solar activity? And how would it compare to changes in the sun's irradiance that are tied to the cycle?

The net sunlight reaching the earth depends on the sun's irradiance and the earth's reflectance. As mentioned earlier, it is generally agreed that a 0.1% variation over the activity cycle is too small to be climatologically significant (see Lean 1997 for a review). What about the earth's reflectance? Taking the aforementioned albedo changes to be real, how significant would they be? To see the relative roles of varying solar irradiance and the varying reflectivity of the earth in the change in the net input of sunlight to the climate system between a time of high solar activity (1999-2001) and one of low activity (1994-1995), we use Eqs. (1) to determine

$$\frac{\delta P_{in}}{P_{in}} = \frac{\delta C}{C} + \frac{-\delta A^*}{1 - A^*}, \tag{6}$$

where we assume the Bond albedo varies like A^*. From this, we note that $\frac{\delta C}{C}$ is 0.1% from solar maximum to solar minimum. Our observations of the earthshine take the ratio of the earthshine to moonshine, so they are insensitive variations of the solar irradiance. The 5%± 2% change in our observed reflectance translates to $\frac{-\delta A^*}{1-A^*} \sim 0.021 \pm 0.007$. Solar and terrestrial changes are in phase and contribute to a greater power going into the climate system at activity maximum. However, the effect of the albedo is more than an order of magnitude greater. Still, whether the earth's reflectance varies with the solar cycle is a matter of controversy, but regardless of its origin, if it were real, such a change in the net sunlight reaching the earth would be very significant for the climate system.

It is always possible that the change derived from the observations have been overestimated, and so we look to a different indicator of this change, which can be derived from the simulations. In this case, we only have the simulations coming from the monthly mean cloud cover from the ISCCP D2 data, and we find a -0.005±0.003 decrease in albedo from 1994/95 to 1999 (1.6%± 0.8%), and so for this period $\frac{-\delta A}{1-A} \sim 0.007 \pm 0.004$. Comparing with the solar irradiance change during the same period, again the two contributions have the same sign, but the effect of the albedo is several times larger.

During the restricted period of 1984-1991, the total and low cloudiness variations from ISCCP were strongly related (see Pallé et al. 2003). Following Willson & Hudson (1991), we compare the simulated albedo changes between 1986 and 1990. These two times correspond to the solar activity minimum and a time near maximum. Actually, 1991 was the maximum, but we chose 1990 because 1986 and 1990 are roughly at the same phase of the El Niño and Mt. Pinatubo erupted in June of 1991. Over the four years between 1986 and 1990, the solar irradiance increased by about 0.1%, while the simulated albedo decreased by about -0.004±0.005 (1.3%± 1.6%). This translates to a $\frac{-\delta A}{1-A} \sim 0.006 \pm 0.007$, roughly the same as for the period 1994 to 1999.

The same trend in the albedo holds from 1990 to 1994, and with about the same magnitude ($\delta A = -0.003 \pm 0.002$ or $\frac{-\delta A}{1-A} \sim 0.004 \pm 0.003$). This behavior is opposite to what one would demand from a solar cycle dependence. However, as we have seen there is the very real possibility that the treatment of the clouds is so inadequate that a correct treatment could change the trends, e.g., from a steady decline in albedo, like in Figure 4, to an oscillating behavior like the IR low cloud cover displays. But, we still don't know.

If the latter hypothesis is to be true and there is a solar influence on the earth's albedo, one question remains: would these changes be climatologically significant over the solar cycle? Looking at our simulations, between 1994 and 1999, they imply an effective increase in the solar irradiance of 10.92±5.96 W/m^2, or a surface averaged (i.e., divide by four) increase of 2.73±1.36 W/m^2 over the five year span, or 7.5± 2.4 W/m^2, as derived from the observations. We change here to a surface-averaged value, since that is the way in which the climate models are presented. Between 1990 and 1994, the change is 1.02±1.02 W/m^2 (where during this period $\frac{\delta C}{C}$ is negative), but between 1986 and 1990, data imply a surface-averaged increase of 2.38±2.38 W/m^2. The latest Intergovernmental Panel on Climate Change (IPCC 1995) argues for a comparably sized 2.4 W/m^2 increase in forcing, which is attributed to greenhouse gas forcing since 1850. IPCC also argues that this forcing is larger than changes in the sun's irradiance, aerosols and their effect on clouds, and that greenhouse gasses have dominated climate change since the industrial revolution being ultimately responsible for the increase in global temperatures experienced during the major part of the 20th century.

The change arising from the albedo, over just five years is too abrupt to change the global climate – considering the thermal inertia of the oceans. However, Stevens & North (1996) have used ocean surface temperature data to suggest a subtle variation, with an eleven year period, since 1850. One key to the argument comes from the study of ^{10}Be concentration in Greenland ice cores by Beer, Raisbeck & Yiou (1991), to show that the cosmic ray flux at activity maximum a hundred years ago was comparable to what it is now at activity minimum. ^{10}Be is created by cosmic rays and falls to earth in about two years. As such, it is a good proxy for the cosmic ray flux.

Relating these changes in radiative flux to changes in the earth's surface temperature is problematic. We focus here on changes in the earth's emission temperature, as would be seen in the IR near 15.5 μm. In those terms, we have from Eq. (2),

$$\frac{\delta P_{\rm in}}{P_{\rm in}} = \frac{4\delta T_e}{T_e}, \tag{7}$$

under the assumption the ϵ in Eq. (2) doesn't change. Taking $T_e \approx 255K$, we find a temperature perturbation due to the sun of about 0.1 K from the irradiance changes, but about 0.5 K and 1.3 K (simulations and observations, respectively) from the albedo. We note

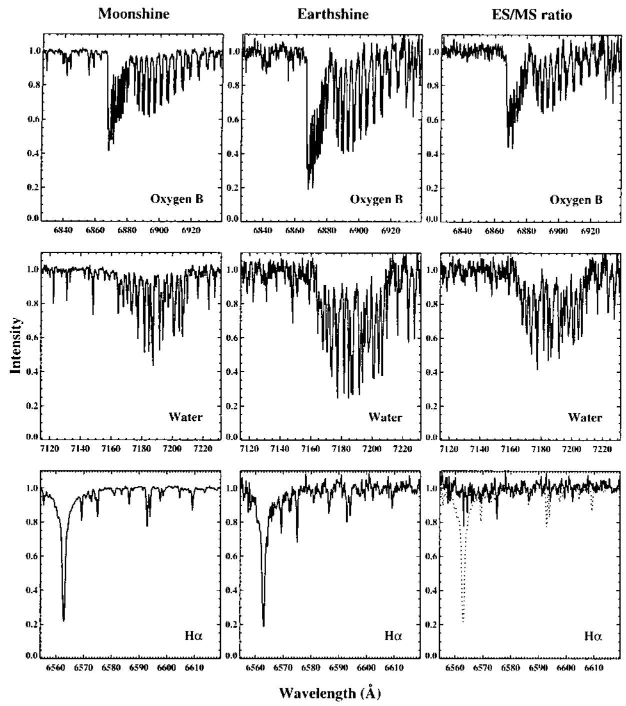

Fig. 5.— The spectrum of the moonshine, earthshine and the ratio of the two in the spectral regions of Oxygen A (top row), water (middle row) and Hα (bottom row).

that the largest variations in Figure 4 correspond to temperature variations of about 1 K for the earth. The temperature changes here are simply related to changes in the earth's net sunlight, not changes in the temperature of the earth's surface.

VIII. THE SPECTRUM OF THE EARTHSHINE

We have been measuring the spectrum of the earthshine from Mt. Palomar using the 60″ telescope, and its Echelle spectrograph. These observations give the only large-scale measure of the earth's spectrum and its variation. These data hold the promise of enabling us to determine the large-scale variation of the structure of the terrestrial atmosphere as well as the evolution of its content of greenhouse gasses. Some typical spectra are shown in Figure 5.

The figure shows partial spectra of the moonshine, earthshine and and the ratio of the two. By taking the ratio of the two, we can eliminate the solar signal (see how Hα is gone from the lower right panel) and the signal of the last pass through the atmosphere above Palomar. The ratio leaves us the large-scale spectrum of the other side of the earth. To analyze these data, we first focus on the moonshine data.

The moonshine data contain the solar spectrum and the spectrum from the clear sky between the telescope and the moon. We carefully fit the terrestrial part of the moonshine spectrum and cross-check it against standard models of the atmosphere. The good agreement gives us confidence that we can correctly fit the earthshine spectrum, where we can't compare with standard models because of complicating factor of cloud cover.

ACKNOWLEDGEMENTS

This work has been supported by NASA grant NAG5-11007

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