ANTHROPOGENIC IMPACT ON THE ALBEDO OF THE EARTH

J. OTTERMAN
Tel Aviv University, Dept of Geophysics and Planetary Sciences, Ramat Aviv, Israel

Abstract. The impact of man and animal on the Earth's surface albedo (reflectivity), until recently believed to be quite small, or not considered at all, is analyzed. Discussion is presented of changes in the albedo due to the heat island effects of cities on snow cover, to agricultural cultivation, irrigation, and to overgrazing; the latter of which is emphasized. In arid climates, protected steppe areas have a low albedo due to dark plant debris accumulating on the crusted soil surface, whereas the same type of terrain, when overgrazed, exhibits a high albedo of trampled, crumbled soil. Extrapolating from observed spatial differences between overgrazed terrain and natural steppe, it is suggested that anthropogenic pressures mainly due to overgrazing could have had a very significant effect on the Earth's surface albedo both regionally and as a global average during the last few thousand years. The Earth's surface albedo presently might be 0.154 whereas it might have been 0.141 about 6000 years B.P. Thus, the surface albedo could have increased by $\Delta a = 0.013$, or by nearly 10% of its value when steppe areas were in their 'virgin' state. The hypothesized increase would be much larger in the Northern hemisphere than in the Southern. There are uncertainties even in the second digit of the suggested value for the present day albedo, and thus certainly for the albedo in the past. Seasonal mapping of the surface albedo from LANDSAT type satellites is recommended.

Behold, the camels were coming.
Genesis, Chapter XXIV, Verse 63

1. Introduction

1.1. Surface Albedo and Its Climatic Significance

Albedo of the surface is the ratio of solar irradiation reflected from the ground to the impinging irradiation on the surface. Thus one minus albedo is the fraction of the irradiation absorbed by the ground. Albedo is closely related to reflectivity, which is usually specified in a relatively narrow spectral band and into a specific direction from the surface. Albedo, then, can be thought of as effective reflectivity, integrated over the hemisphere[1] of all possible reflection directions and over all the wavelengths of the solar irradiation.

The values of the albedo vary greatly with type of terrain. Averaged values of albedo are shown in Table I, from Budyko (1974; p. 55). For water surfaces, the albedo is quite low, 0.06 (except that it is higher for high sea-state or very low sun illumination angle).

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TABLE I: Mean values of albedo (from Budyko, 1974)

Type of surface		Albedo
Stable snow cover in high latitudes (more than 60°)		0.80
Stable snow cover in middle latitudes (less than 60°)		0.70
Forest with stable snow cover		0.45
Unstable snow cover in spring		0.38
Forest with unstable snow cover in spring		0.25
Unstable snow cover in autumn		0.50
Forest with unstable snow cover in autumn		0.30
Steppe and forest during the transition period after snow cover disappears and before the mean daily temperature of the air reaches 10°		0.13
Tundra without snow cover		0.18
Steppe or deciduous forest during the period after the temperature goes above 10° in spring and before snow cover appears in autumn		0.18
Needleleaf forest during the period after the temperature goes above 10° in spring and before snow cover appears in autumn		0.14
Forest dropping leaves during a dry season, savanna and semidesert	During the dry season	0.24
	During the wet season	0.18
Desert		0.28

The significance of surface albedo from the climatogenic point of view lies in the first place in the fact that there is a strong correlation between the surface albedo of an area and the albedo of the Earth-atmosphere system, as seen from space above it. The albedo of the Earth-atmosphere system determines directly how much of the solar irradiation is reflected into space, and how much is absorbed by our climatic system. For instance, from recent measurements of Earth-atmosphere system albedo over ice-covered Greenland, very high measurements of 0.80 were reported (Jacobowitz, 1976), compared with the average Earth-atmosphere albedo of about 0.30 for the Earth as a whole. Even though, for instance, the Earth-atmosphere albedo above the dark oceans is several times higher than 0.06 of the surface (depending largely on the cloud statistics in the region); the surface albedo still determines in a large measure the intake by the Earth of the solar radiation.

A related significance of the surface albedo lies in the question what is the ratio of the amount of the solar radiation absorbed by the surface to that absorbed by the atmosphere. A fairly recent estimate was that, in terms of the solar irradiation at the top of the atmosphere, 51% is absorbed by the surface and only 19% are absorbed by the atmospheric water vapor, CO_2, ozone, dust and clouds (U.S. Committee for GARP, 1975, p. 18). Thus, the ratio of energy absorbed by the surface to that by the atmosphere appears to be about 2.5 to 1. A large fraction of the energy absorbed by the surface (about 60%) is supplied to the atmosphere in the form of latent heat of vaporization and direct thermal (i.e., sensible heat) flux. The input of sensible heat, and in a different way the input of latent heat, cause convective motions thus 'driving' the atmosphere. It is an open question whether this important ratio has been accurately estimated.

The third significant aspect of the surface albedo is its spatial variability. The heating rates of the surface are different, depending on the solar elevation, i.e., time of day, latitude and season. In addition, the wide variation in the albedo values among open land, open ocean, and ice/snow areas causes additional heating rate differences resulting in spatially variable input of sensible and latent heat into the atmosphere and the oceans. These spatially variable inputs of heat are the basic cause of oceanic and atmospheric circulation. In these phenomena other properties of the surface, such as thermal inertia, might be equally significant or even overriding the albedo differences (consider the case of the sea breeze, which is linked to the higher daytime temperatures of land as compared with the sea), but the albedo differences are still definitely important.

1.2. Measuring the Surface Albedo and Its Temporal Variability

While accurate measurements of albedo are a rather complex task, the different basic approaches to the measurements can be stated quite simply. Measurements can be taken in a laboratory or in-situ from the ground, aircraft or satellite. The most direct approach is to measure the intensity of the impinging radiation, by directing a radiometric instrument with an hemispheric, i.e., 2π radians, opening upwards, and then directing the instrument downwards to measure the hemispheric reflected radiation. This approach is difficult in a laboratory, because it involves preparing a large sample area. For field measurement, it is a simple approach for a homogeneous terrain, but when the surface is inhomogeneous (for instance plants and bare ground), it is difficult to select a representative sample.

In the laboratory a calibrated irradiating beam is used; then integration is carried out of measured spectral reflectivities over the solar spectrum at the ground computed for the conditions of interest.

A different approach is to measure the reflected radiation in parts, separately in each direction, using a radiometer with a narrow beam opening. Each individual measurement is often called in such case the bi-directional reflectivity, since it depends on the elevation angle from the surface and the azimuth relative to the irradiating beam of the reflected beam. These individual measurements are then summed up for the whole hemisphere.

Still another approach is based on measuring the reflectivity in just one direction. This measured quantity is usually called beam albedo, rather than bi-directional reflectivity, when the surface irradiation is the actual solar irradiation and not an artificial beam. Subsequently the beam albedo is converted into a hemispheric albedo, assuming a simplified angular distribution of the reflection into a whole hemisphere.

The last approach was used by Kung *et al.* (1964) in a low level aircraft flights program of albedo measurement over the North American continent. In such low level flights the atmospheric effects can be neglected. Kondratyev *et al.* (1974) pointed out that for values of ground albedo in the range of 0.20 to 0.50, the Earth-atmosphere albedo when measured from space (in the absence of clouds) differs only slightly from the surface albedo. Otterman and Fraser (1976) computed the correction for the atmospheric effects necessary for calculating the surface reflectivity from the space-measured reflectivity, and subsequently used data from the LANDSAT Multispectral Scanner digital tapes to compute the reflectivities and the albedo in some selected arid regions. Since the LANDSAT Multispectral Scanner measures radiances in the upward vertical direction from a small area on the ground, the approach is basically also a beam albedo measurement. Otterman and Fraser (1976) found the albedo of sandy soils ranging from 0.34 to 0.52. These values are much higher than 0.28 of deserts suggested from Table I. There are thus considerable uncertainties in the measurements, which can be only in part explained as due to the temporal variability of surface albedo. However, Otterman's and Fraser's (1976) results seem to be strongly supported by older laboratory measurements for various sands (Hovis, 1966; Gates, 1970), by ground measurements of terrain components using a hand held radiometer, reported in Table II and discussed later, and by the very recent report of Nimbus 6 results that the space albedo of Sahara at 25° N 13° E was found to be 0.38 in the absence of clouds, (H. Jacobowitz, 1976).

The variability of surface albedo arises from several causes. The character of the surface can undergo changes depending on time of day, season or meteorological conditions. White-caps, for instance, sharply increase the albedo of water bodies.

A systematic variation occurs with time of day, because of geometrical effects. The albedo of many surfaces increases at low elevations of the irradiating beam from the horizontal. Thus there is an increase in albedo at low solar altitudes. This dependence becomes less pronounced under cloudy conditions, because an increase in cloudiness reduces the direct fraction and increases the scattered fraction of solar radiation, and the absorption of scattered radiation depends little on the altitude of the sun (Budyko, 1974; pp. 54–55, referring to earlier work of L. A. Biriukova).

Still another cause of the variability lies in unequal reflectivity from one band to another within the solar spectrum and the changes in the spectral distribution of ground

TABLE II: Spectral reflectivity measurements by Exotech-100 LANDSAT ground-truth radiometer.

	MSS band				
	4 0.5–0.6 mμ	5 0.6–0.7 mμ	6 0.7–0.8 mμ	7 0.8–1.1 mμ	Estimated surface albedo
Crumbled Sinai soil No. 1	0.32	0.44	0.50	0.54	0.46
Plant debris	0.09	0.13	0.17	0.26	0.18
Sinai soil No. 1 crust	0.28	0.35	0.39	0.43	0.37
Artemisia monosperma young growth	0.30	0.20	0.47	0.56	0.38
Artemisia monosperma twigs	0.29	0.21	0.31	0.43	0.32

irradiation as a function of air mass traversed. For three important classes of the surface – soils, vegetation and debris (dead plants littering the surface) – the solar infrared reflectivity is much higher than the visible band reflectivity. The solar infrared irradiation constitutes somewhat less than 50% of the total incident solar irradiation at the top of the atmosphere and can be more than 50% at the ground, depending on the solar altitude and the water vapor levels. For 20 mm precipitable water vapor, the infrared fraction is 50% when the solar beam transverses one airmass, i.e., the sun is at the zenith. The infrared fraction increases with an increasing solar zenith angle, to 61% at 5 air masses (Smithsonian Meteorological Tables, 1958, p. 438). For these three classes of the surface, the morning albedo can be appreciably higher than the noon albedo, and the effect changes with latitude, season and meteorological conditions (Otterman and Fraser, 1976).

1.3. Anthropogenic Influences

The anthropogenic modification of the Earth's surface had been studied by geographers already in the nineteenth century (Marsh, 1864) and attracted considerable attention in the twentieth one (e.g., Thomas, 1956; SCEP, 1971). From the considerations of climate, the modifications of the water cycle was stressed. This viewpoint is quite appropriate, since a water-rich surface in an arid region can produce an evaporation or evapotranspiration of more than 1000 mm per year. Thus irrigation reservoirs, fish ponds and irrigated fields do indeed produce a significant input of water vapor to the atmosphere, while cutting down of forests and jungle can significantly reduce this input.

The effects on the surface albedo were considered small (SMIC, 1971), but perhaps deserving further study. A possibility was raised that an oil spill in the Arctic ice-pack region might cause a sharp regional reduction of the surface albedo, possibly with irreversible climatic effects (Campbell and Martin, 1973).

Specifically regarding the effects of grazing by cattle, goats, sheep or camels, it was recognized that wide areas are subject to this anthropogenic pressure (Curtis, 1956). But as actually measured in areas of moderate or high precipitation (Howard, 1971), overgrazing was thought to decrease the albedo slightly. The unexpected finding from satellite imagery is that overgrazing in arid and semi-arid regions has a sharp impact of increasing the albedo, as discussed later. While perhaps unexpected, this finding should not be regarded as surprising. Under moderate or high precipitation, natural vegetation covers 100% of the terrain in multiple layers of leaves, and the direct effect of grazing is a reduction of the plant canopy. In the semi-arid regions, moisture-limited vegetation covers only a fraction of the terrain even when not hindered by man or animal. Each plant can then gather moisture from the larger surrounding area, which is devoid of vegetation. Dark plant debris tend to accumulate in a thin layer on these areas. Grazing and overgrazing does reduce the plant cover fraction, but more observable and important effects are the impact that the trampling animals have by crumbling the soil on the interstices between the plants. Thus, in this paper processes in plant interstices over the arid regions of the world are underscored. The vast geographical scale of these processes – about 7% of the Earth's surface – justifies the special attention. It is suggested that important changes in the Earth's albedo could have occurred from the prehistoric times, as a result of the anthropogenic impact.

2. Artificial Water Bodies, Irrigation, Agricultural Cultivation, Fire and Heat Island Effects on Snow

The albedo of clear water bodies is quite low, about 0.06. For the darkest soils, the albedo can be in the same range, but for the bright, sandy soils or sand, it can range up above 0.50. Thus creation of artificial water bodies can result in a sharp decrease of the albedo. In Figure 1, LANDSAT image of Lake Nasser in Egypt, the dark lake appears in a sharp contrast to the surrounding mostly sandy, desert. Note however that in the generally bright desert there are also pronounced variation of reflectivity.

Moist soils have a lower reflectivity than dry soils, typically by a factor of two (for bright sandy soils) throughout the visible and the reflective infrared wavelengths. Irrigation, therefore results in a very pronounced lowering of the albedo of the soil.

Fig. 1. LANDSAT image of Lake Nasser in Egypt (lower right corner) and the surrounding desert, 185 x 185 km.

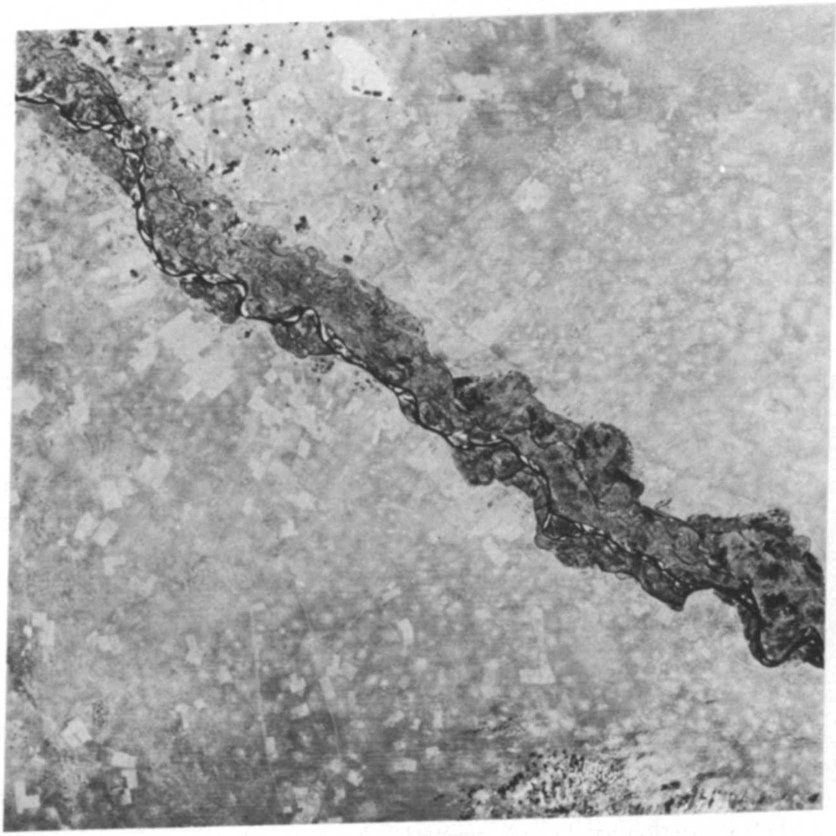

Fig. 2. Semi-arid region of the lower Volta. Dense vegetative cover near the river appears dark in the image, but has high infrared reflectivity. Large fields are visible, mostly as bright rectangles on both sides of the river. Many settlements in the image, especially on right, are surrounded by bright circular 'haloes' of about 4–5 km, indicating the range of grazing by sheep, 185 x 185 km.

The real comparison can be made only by comparing an actual irrigated and cultivated field with the terrain and its natural vegetation as it appeared before the irrigation started. Because of its high infrared reflectivity (Coulson *et al.*, 1965; see also Table II), albedo of green vegetation and crops when forming a solid canopy is quite appreciable, typically in the range 0.20 to 0.40, in spite of the dark appearance in the visible. The question of what impact the agricultural cultivation has on the albedo is important, but in each area it depends on the type of soil, natural vegetative cover which existed before the cultivation, types of crops and the agricultural practices (Figure 2). The albedo of crops and vegetation is seasonally variable, and that includes bare soils because of the changing moisture contents. No general assessment, even whether the cultivation increases or reduces the albedo, can be presented. Regional mapping of the surface albedo, on a seasonal basis, from LANDSAT type satellites, is required to gather information on the subject.

Man-induced fire is certainly a powerful mechanism by which man even in the early civilizations could modify a relatively large area for his purpose in his agriculture-related activities. Today, in Dahomey, for instance, the use of fire for agricultural purposes is a standard practice, and it can be said that a large fraction of Africa is burned annually (McLeod, 1976). The dark appearance of the burnt-out areas wears out in about a month under the influences of weather and cultivation, especially ploughing (McLeod, 1976), and thus the darkening of the surface is a limited seasonal effect. The induced soil instability due to plant eradication is much more important, and has to be added as a serious contributing agent along with overgrazing, see the next section.

Heat-island effects of cities or other man-made installations can seasonally produce very dramatic effects on albedo, by their snow melting action. The resulting decrease of the surface albedo, from up to 0.80 of fresh snow (Kondratyev, 1969, p. 417), is very sharp, but only of relatively brief duration, in places and times where the regional solar irradiation is far from its peak values. The LANDSAT images of Leningrad, Moscow and Kharkov in February or March, show each city as a big circle free of snow, whereas patchy snow covers the countryside.

3. Effects of Overgrazing on Land Surface

The characteristics of vegetation are high reflectivity in the reflective infrared and low reflectivity in the visible. For rangeland plants the reflectivity in most of the reflective infrared region of the spectrum typically can be between 0.35 to 0.70. In the visible, the reflectivity typically lies between 0.08 and 0.25 only. In the orange-red light especially, where a chlorophyll absorption band exists, vegetation is quite dark.

The infrared reflectivity of a plant canopy increases with an increase in the percentage of ground cover and in the canopy thickness, i.e., number of layers of leaves (Vinogradov, 1969; Colwell, 1974). These well known reflectance characteristics offer means for remote sensing of biomass of rangelands from the LANDSAT satellite: by mapping the ratio of the reflectivity in the infrared 0.8–1.1 μm band (called MSS-7 in LANDSAT) to the reflectivity in the orange-red band, 0.5–0.6 μm (called band MSS-5). This approach was proven to work well in mapping the rangeland biomass in the U.S.A. (Rouse *et al.*, 1973).

Entirely consistent with these reflectance properties were measurements of Howard (1971) in Australia that grazing produces a slight reduction in the albedo. More specifically, Duggin *et al.*, (1975) studied the effects of controlled grazing on the spectral reflectivity, by the EXOTECH 100 ground observation radiometer operating in the LANDSAT Multispectral Scanner bands. They found that grazing resulted in an increase of the reflectivity in the visible bands (difference of 2 to 3%). This brightening was, however, more than offset by a decrease of the reflectivity in the infrared bands (difference of 3 to 7%), confirming the over-all darkening effect, i.e. reduction in the albedo, reported by Howard. In the ungrazed paddocks, lucerne covered essentially 100% of the ground. The annual rainfall in the test area is 800 mm (Duggin, 1975).

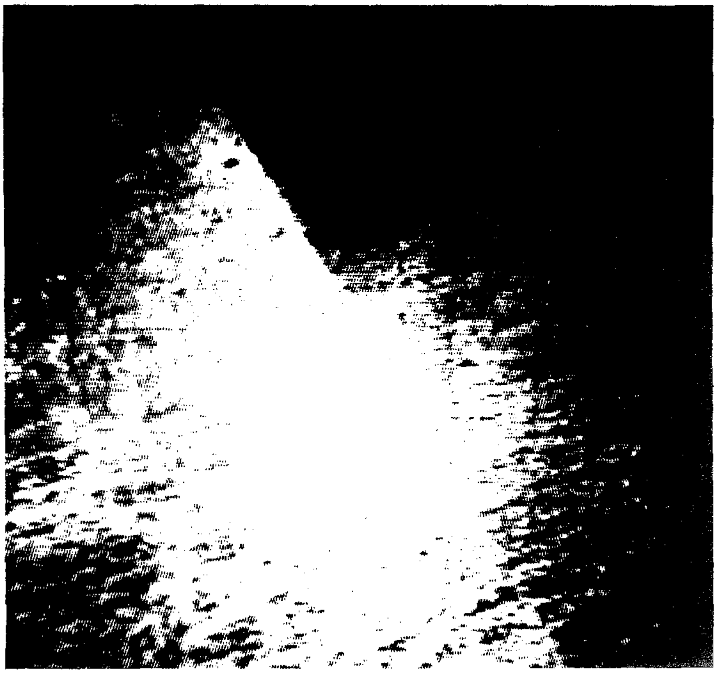

Fig. 3. Negev/Sinai demarcation line, a photograph of high-resolution display from LANDSAT digital data. Dark Negev in upper right.

The operation of LANDSAT satellite over Israel provided excellent means of studying the effects of overgrazing. Sinai and Gaza strip have been continually overgrazed and subject to additional anthropogenic pressures: picking up of plants for firewood and for construction of desert habitats by the Bedouin. There is also cultivation, especially of almonds and watermelons, in small plots. In contrast, since the 1948/9 armistice, the Israeli Negev (to the Northeast of the Sinai and East of the Gaza Strip) has been by and large free from such pressures, and the natural vegetation grows more abundantly than in the Sinai. The Sinai/Negev demarcation line has been observed in a sharp contrast by Lowman (1966), in a photograph from a manned spacecraft, i.e., an image in the visible part of spectrum (Lowman attributed mistakenly the contrast to cultivation in the Negev). The LANDSAT imagery, starting with the first pass October 22, 1972, provided high resolution imagery of the area (see Figure 3) also in the reflective infrared part in the spectrum, and the surprising observation was that the sharp demarcation line, at roughly the same contrast as in the visible, was observed also in the infrared bands of the LANDSAT, MSS-6 (0.7–0.8 μm) and MSS-7 (0.8–1.1 μm) (Otterman, 1973).

It should be pointed out that a considerable spread exists in reflectance characteristics of various soils and various plants and that some desert sands *can* be brighter, i.e., more reflecting in the infrared than some plants (Gates, 1970). However, this could not be offered as an explanation, since the sharpest contrast was observed in the sandy loess area, rather than in the brightest sand dune areas, and since the contrasts were about the same in all the bands.

The plants in such arid zones as the Negev are moisture limited and cover only a fraction of the terrain – 25% in the Negev and 10% in the Sinai can be taken as representative fractions (in the areas where the annual rainfall is about 150 mm/year). The offered explanation lies in the characteristics of the plant interstices, which constitute the bulk of the area. The important role of the vegetation in the Negev is to stabilize the soil – a crust forms, and dark plant debris litter the interstices. In the Sinai, trampled by the grazing herds, the bare and crumbled soil shows up bright compared to the debris-covered soil in the Negev (Otterman *et al.*, 1975).

The brightening of the areas unstabilized by grazing and anthropogenic pressures are by no means unique to the Sinai/Negev demarcation line, and indeed, have been found in many parts of the world, where across a political boundary different cultivation and settlement patterns exist. Such a contrast across U.S.S.R./Afghanistan border can be seen in Figure 4. A similar effect with a lower contrast, extends for hundreds of kilometers in Southwest Africa in a straight line across the Namibia/Botswana border. But the highest

Fig. 4. Afghanistan/USSR boundary, thematic mapping of overgrazed areas rendered in white.

contrast, about 2.0, was measured in a section of the Negev/Sinai demarcation line (Otterman *et al.*, 1976).

The albedo of overgrazed soils and unstable sands cannot be specified narrowly. A rather wide range of values from 0.34 to 0.52 was found in various arid and semi-arid locations by radiometric analysis of LANDSAT imagery (Otterman and Fraser, 1976). These values are higher than most researchers reported from space, but correspond quite well to the ground truth measurements.

In-situ reflectivity measurements of terrain components were recently carried out in the Sinai, near where Sinai, Negev and Gaza Strip all join at one point, by EXOTECH-100 radiometer, and are presented in Table II. Weighed averaging of band reflectivities yield for the crumbled Sinai soil reflectivity of 0.40 in the visible and 0.53 in the infrared, for an estimated albedo of 0.46. This value checks rather closely with the LANDSAT measurements in the same vicinity for the Sinai terrain (Otterman and Fraser, 1976).

The crusted soil in the Negev was typically found in the EXOTECH-100 measurements by 10% or 20% darker than the Sinai crumbled soil. Soil covered solidly with plant debris was darker than the crumbled soil by a factor of 3.5 in the visible (reflectivity 0.12) and a factor of 2.2 in the infrared (reflectivity 0.24), for an estimated albedo of 0.18.

The grazing herds produce in situ regions of bright, unstable soil by reducing the stabilizing vegetation and by trampling the crust of the stable soil and burying the debris. But the unstable soil can be transported then by the winds, giving rise to bright eolian deposits at some distance from the overgrazed area. Pronounced eolian formations have been observed from space imagery by Lowman (1971), who stressed the importance of such processes.

4. Possible Global Impact of Overgrazing and Other Anthropogenic Pressures

4.1. Earth's Albedo at the Present

As will be seen from the discussion which follows, there are very large uncertainties in the present day surface albedo of the Earth.

Flohn (1975) presents a schematic estimate of the surface albedo, just for the study of the effects of shifts in the ice cover of the Earth. His computations are reproduced in Table III.

Flohn assumes 0.12 as the albedo for the ice-free land areas. This seems a rather low value even for other areas than unstable, sandy soils, in view of the generally high reflectivity of vegetation in the infrared (Coulson *et al.*, 1965). This low value is particularly inappropriate for sandy deserts and overgrazed areas of unstable soils for which 0.42 can be taken as the average albedo in view of the measurements by Otterman and Fraser (1976). Such areas are assumed here to constitute $1/3$ of open, i.e. ice-free and snow-free land areas, as explained below. We retain 0.12 as the albedo of $2/3$ of the open land areas, but take it to be 0.42 for the other $1/3$. A correction is computed then to Flohn's albedo as follows:

For the Northern hemisphere

$$\Delta a = \frac{\frac{1}{3}\text{ ice-free land area} \times (0.42 - 0.12)}{\text{total area of hemisphere}}$$

$$= \frac{23.3 \times 0.3}{255} = 0.0274$$

and for the Southern hemisphere

$$\Delta a = \frac{11 \times 0.30}{255} = 0.0129.$$

Adding this correction of Flohn's values, we obtain 0.1568 as the albedo of the Northern hemisphere, and 0.1513 as the albedo for the Southern hemisphere. It should be noted that the albedo of the Northern hemisphere is here higher than that of the Southern.[2] The difference would have been still higher had we raised the albedo of the ⅔ of ice-free land above 0.12.

The area of sandy deserts and overgrazed rangelands in arid and semi-arid regions was estimated as ⅓ of ice-free land areas, i.e., as 23.3×10^6 km^2 in the Northern hemisphere and as 11×10^6 km^2 in the Southern hemisphere, for a total of 34.3×10^6 km^2. This is only an approximate estimate, and for true consideration of the climatic effects problem, a breakdown of such areas at the various latitude bands should be presented. This was not attempted. The semi-arid, arid and extremely arid regions of the world constitute 36% of the total land areas (Amiran, 1966). This 36% includes stony deserts, which we have to exclude. Other pertinent information is that rangelands and meadows (which includes moist regions but excludes sandy deserts) covered 29×10^6 km^2 in 1968 (FAO Sources, quoted in SCEP, 1970, p. 280).

Still another consideration was connected with the number of sheep and goats in the world, reported as 1030 million sheep and 369 million goats for 1963–5 (Jones, 1972). The number of goats and sheep in the borderline arid semi-arid region of Beersheva is 40 per square kilometer. This is somewhat above the carrying capacity for the region (without other food supplement), and thus results in some overgrazing (Noy-Meir, 1975). Taking 40 units per square kilometer as a global average, for meadows and rangelands, we obtain $(1030 + 369) \times 10^6/40 = 35 \times 10^6$ km^2, very roughly confirming our estimate of 34.3×10^6 km^2, (which includes extremely arid sandy deserts, where grazing is impossible).

4.2 Earth's Albedo 6000 B.P.

The uncertainties in what the albedo was a few thousand years ago are obviously large. Still, a presentation is made of a possible change in the albedo over the last few thousand years. Only a *rough* estimate of the order of magnitude of the *possible* effect can be presented.

What was this area of presently overgrazed rangelands and sandy deserts like some

TABLE III: Surface albedo, present and 6000 B.P.

Flohn (1975)							Modified[a], present			Modified[a], 6000 S.P.		
							Open continents			Open continents		
Albedo	Open 0.05	Ice 0.70	Open 0.12	Ice 0.75	Snow 0.30	Average albedo	2/3 0.12	1/3 0.42	Average albedo	2/3 0.12	1/3 0.22	Average albedo
Northern hemisphere	145	10	70	3	27	0.1294	46.7	23.3	0.1568	46.7	23.3	0.1385
Southern hemisphere	190	16	33	13	3	0.1384	22	11	0.1513	22	11	0.1427
Earth	335	26	103	16	30	0.1339	68.7	34.3	0.1540	68.7	34.3	0.1406

[a]In the modified calculations, Flohn's (1975) values are retained of the surface albedo for the oceans, for the ice and snow covered continents and for 2/3 of the open continents. Areas in 10^6 km^2, after Flohn (1975).

6000 years ago? We hypothesize (with only slight supporting evidence, as discussed later) that the vegetative cover was more ample in the arid regions. We assume here that the climatic change was only slight, i.e., the total area of semi-arid, arid and extremely arid lands was the same as today. 6000 B.P. falls in the middle of the warmest period of the Holocene, our present inter-glacial period (U.S. Committee on the GARP, 1975, p. 40) and of the 'Neolithic wet phase' in Sahara (Huzayyin, 1956; Butzer, 1966). If indeed precipitation at the present day 50 mm isohyets in North Africa was 150 mm (as discussed by Butzer and Twidale, 1956), this would not necessarily negate our climatic assumption.

With this climatic assumption, the 34.3×10^6 km^2 area under discussion would appear darker, just like today's animal exclosures in the arid regions, mainly because of plant debris littering the interstices. The bright sandy soils, solidly covered by the plant debris can have an albedo lower by nearly a factor of 3 than the bare soil, as the recent measurements in the Negev indicate. If 6000 years B.P. the albedo of today's overgrazed rangelands and sandy deserts was 0.22, like that of the darkest area in the Negev, the hemispheric albedos can be estimated as 0.1385 for the Northern hemisphere and 0.1427 for the Southern hemisphere. Our calculated present day albedo is a pronounced departure from these values. The Southern hemisphere was brighter than the Northern, and the Earth as a whole darker than today by $\Delta a = (0.2 \times 34.3)/510 = 0.013$.

5. Discussion and Conclusions

Overgrazing and anthropogenic pressures of collecting shrubs for firewood and for construction of habitats reduces the vegetative cover. In arid regions, where (a) vegetation grows in clumps with large interstices; (b) wind erodibility is potentially great, because of prevailing dry conditions, an important accompanying effect of overgrazing and anthorpogenic pressures is the destabilization of the soil. Soil instability affects the albedo, in two ways: by microscale effects, in producing on the steppe interstices of crumbled soil, free from plant debris cover; and by macroscale effects, by creating areas of bright sand dunes and shifting sands that can cover the steppe and the neighboring areas. The increase in albedo can be from a range of about 0.20–0.28 to a range of 0.34–0.52 and possibly higher for the very recent sand dunes. An increase by a factor of about 2.0 has been measured in a section of Sinai/Negev demarcation line (Otterman *et al.*, 1976).

Surface albedo can be subject to important modification when the anthropogenic pressures are removed, as recently evidenced by a 6 km x 6 km fenced exclosure erected in the Northern Sinai by the Israeli Army in the summer of 1974. The April 1976 LANDSAT imagery already shows the fenced-in area darker, at a high contrast to the surrounding overgrazed terrain.

Extrapolating from the spatial differences that were observed between overgrazed regions and exclosures to the temporal differences, under an assumption that overgrazing and anthropogenic pressures were not prevalent some 6000 years ago, we arrive at a

Fig. 5. Dust storm in the Persian Gulf.

conclusion that a rather dramatic increase in global albedo, by $\Delta a = 0.013$, or about 10% of the mid-Holocene albedo, could have occurred as a result of such processes.

Soil instability over wide areas can give rise to dust storms. Dust storms do not materially affect the surface-atmosphere system albedo when occurring over the deserts, but when the dust is carried to over the ocean, a large increase in albedo occurs temporarily and locally. A processed LANDSAT image of a dust storm, over the Persian Gulf is shown in Figure 5. The albedo (computed as surface albedo assuming a relatively clear atmosphere) in the dust cloud is about 0.29, the albedo of the darker area of the sea is 0.17. These figures should be compared to a typical value of ocean surface albedo of about 0.06.

The suggestion of possible dramatic changes that possibly had occurred is presented to point out the need of further interdisciplinary studies. At the present there are only hints that provide some supportive evidence. Recent studies underscore the importance of aeolian processes during the Quaternary (Lowman, 1971; Yaalon and Ganor, 1973; Jackson *et al.*, 1973). More specifically to Holocene, Yaalon (1975) reports that the recent rate of deposition in some locations in Israel exceed the past mean long-term rates of a few thousand years ago. Current studies by Yaalon and his colleagues of secular changes in paleomagnetism in a thick loess section in the northern Negev might provide much fuller information on this subject (Yaalon, 1976).

Another, possibly more direct, indication of the plausibility of the outlined aeolian processes can be found in the studies in the Negev by Dan (1973). He found in the Northern Negev buried Hamra and related soils, which nowadays characterize the sub-humid Sharon coastal plain of Israel. Moreover, in the mildly arid areas of the northern Negev, subsurface aeolian clays are much finer than the recent uppermost loessial soils. As a rule, the comparatively coarse aeolian sediments settle out near the desert, while the finer ones are carried further away. Dan takes the subsurface fine aeolian clays as an evidence that the desert has spread since the late Pleistocene.

Palynological studies (fossil pollen analyses) by Horowitz (1973, 1976) indicate that the general area of Israel was more richly covered with vegetation 6000 years ago than recently. However, this information pertains to the extent of the forest and not to the vegetation of the steppe – there are no comparative pollen counts of the buried fine aeolian clays in the Negev versus the coarse surface deposits.

Selecting 6000 B.P. as the time from which anthropogenic pressures began to mount was influenced by estimates of the world-wide population and also by regional studies in Israel by Gophna (1973). He reported that the Besor area of the Northern Negev was very densely populated in the Chalcolitic age 3400–3100 B.C. but there are virtually no archeological indication of prior settlers. Worldwide, it has been estimated that the human population was under 20 000 000 (Ehrlich and Ehrlich, 1972). Population density of man and his goats must have been well below the carrying capacity of the land, too low to result in appreciable anthropogenic-related pressures.

In the last 6000 years the sea level was essentially stable, with oscillations and secular changes of the order of few meters only, and thus the changes in the extent of the sea did not affect the Earth albedo very markedly.

The studies cited here and studies from other areas point out that our planet *could* have been significantly altered by anthropogenic pressures. Further studies in the fields of archeology, sedimentology, palynology and climatology are expected to provide more substantive answers. There always remain uncertainties. It can be pointed out that aeolian deflation ultimately results in formation of a desert pavement, which can be quite dark. What is today the total area of the dark pavement and how large was it 6000 B.P.? The observed spatial contrasts are between overgrazed areas and partially protected areas, on which grazing was limited for a few decades or a few years. The ecosystems in the protected areas did not necessarily revert to their original state. The high contrasts essentially stem from the plant debris covering about half of the surface. Were the interstices of the original vegetation as large as today? Further questions can be raised, such as: Are the changes in vegetative cover and by aeolian erosion due directly to the anthropogenic pressures or to a climate change or both? If a climatic change did take place during the last 6000 years, was it caused by the anthropogenic pressures? At least three different theories of regional desertification have been presented in the last decade (Bryson and Baerreis, 1967; Otterman, 1974; Charney, 1975) and a strong case made by Flohn (1974) linking global climatic changes to the surface albedo changes.

A multi-disciplinary study to quantitatively reconstruct the surface of the ice-age Earth 18 000 B.P. has been recently conducted (CLIMAP Project Members, 1976). A similar study concerned with the surface 6000 B.P. seems worthwhile, even though we are unlikely ever to know with assurance what the surface albedo of the Earth was 6000 years ago. A surprising point which emerges from recent studies is that as large uncertainties exist today regarding the present surface albedo as that of ages long past. Apart from the studies by Flohn, other workers, such as Kukla and Kukla (1974), stressed the climatogenic importance of the surface albedo changes. A consensus exists that the surface albedo (as a function of the latitude) is an important climatic parameter (GARP JOC, 1975, p. 81). The satellite technology currently makes seasonal monitoring and mapping of surface albedo in cloud-free areas possible, and such projects should be undertaken.

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Notes

[1] The term 'hemispheric albedo' is sometimes used.
[2] The prevalent opinion is that the Southern hemisphere has a higher surface albedo.

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