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Spring Snow Dissipation In Alaska

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ABSTRACT

A regionally parameterized value of absorbed solar radiation at the earth's surface of approximately 260 cal cm^{-2} day^{-1} was reached immediately prior to periods of rapid snowmelt and sharply rising temperature in the springs of 1979-1981 over forested and tundra sections of Alaska. This threshold value, which is a function of solar insolation, estimated surface albedo, and a constant atmospheric screening factor, occurred on the average 42 days later on the North Slope than in the Tanana and upper Yukon River valleys due to differences in regional albedo ranging up to 45 per cent when fully snow covered. Intraregional differences in timing were due to snowpack conditioning prior to threshold attainment, such as advection of warm air into a region or a snowfall event which, respectively, advance or retard the date by subtly altering surface albedo. High temperatures in most cases remained below 5°C prior to threshold attainment, 5-13°C during rapid snowmelt, and over 13°C following melt. Results are based on analyses of satellite-image brightness over Alaska and ground-station data during 1979-1981. A potential CO_2-induced warming in Alaska could result in significantly earlier threshold attainment if snow depth remains within the range of present-day values. However, an increase in the snowpack as a result of warmer, wetter winters could act to delay the threshold.

INTRODUCTION

Snow cover is a critical variable in the climate system, particularly in the high latitudes where the ground may be snow covered from 6 to 9 months each year. Snow can have an impact on microclimate, synoptic weather, and atmospheric circulation (Weller et al. 1972; Dewey 1977; Kukla 1981; Walsh et al. 1981; McFadden and Ragotzkie 1967; Namias 1962, 1980).

Climate models have shown the snow-albedo-temperature feedback mechanism to be a critical factor in the climate system (Sellers 1969; Budyko 1969; Manabe and Wetherald 1975, 1980). Models also indicate that increases in carbon dioxide will lead to an enhanced rise of high-latitude, spring surface temperature resulting from the early dissipation of snow cover (Ramanathan et al. 1979).

We have found a strong association between temperature and surface albedo during several springs in Alaska to be a partial function of the amount of energy regionally available for surface heating, which in spring is primarily controlled by the extent of snow cover.

DATA

Advanced Very High Resolution Radiometer imagery from National Oceanic and Atmospheric Administration polar orbiting satellites was examined for the springs of 1979-1981. Sensors record radiation in wavelengths from .6-.7μm and have nadir resolution

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of 1.1 km. Reports of surface air temperature, precipitation, snowfall, and snow depth are from Climatological Data (NOAA).

PROCEDURE

An image processor was used to measure the surface brightness of select regions of Alaska, including two predominantly forested areas, the Tanana and upper Yukon valleys, and a portion of the tundra-covered North Slope (fig. 1). The processor treats an image as 307,000 discrete pixels, assigning each a gray-scale value from black (0) to white (255).

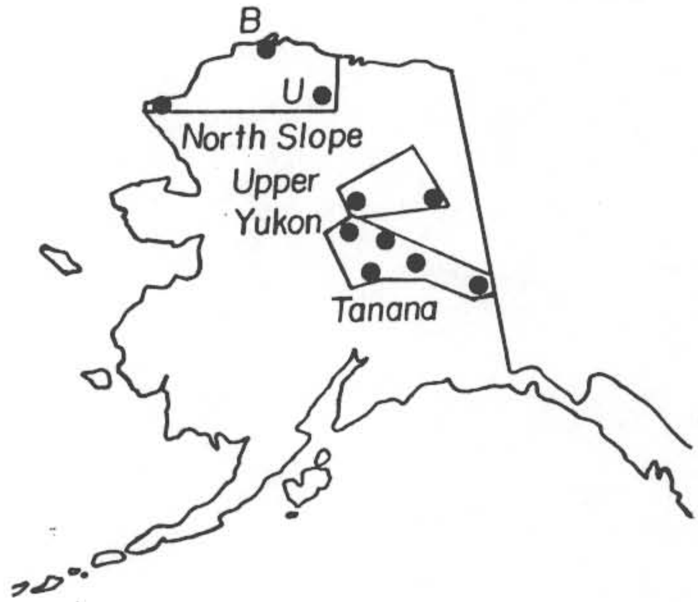

Figure 1. Locations of study regions and ground stations. (U-Umiat, B-Barrow)

Correlation of brightness between images was achieved by setting portions of each image where brightness was assumed constant to a fixed brightness through aperature adjustment of the processor lens (e.g. tundra in early spring, snow-free forests in late spring). This technique minimizes differences in brightness between images caused by atmospheric variability.

Adjusted brightnesses were converted to estimated surface albedos by linear interpolation between a bright, snow-covered surface with a parameterized albedo of 80 and dark, vegetated, snow-free land with an albedo of 13. Regional surface albedo under full snowcover ranged from 35 per cent in the Tanana Valley to 80 per cent on the North Slope. Regional climatic data were derived by averaging reports from stations evenly distributed within the areas (fig. 1).

RESULTS

An example of the close association between sharply decreasing surface albedo and rapidly rising temperatures which was seen in all regions in each of the three springs is shown in Figure 2. Note that:

1. All averages remain below 0°C prior to the commencement of melt (decreasing albedo), with the extreme high only reaching 5°C.
2. Highs during snowmelt average 10°C while lows remain below freezing.
3. Highs average 20°C and lows above 0°C immediately after the melt is completed.

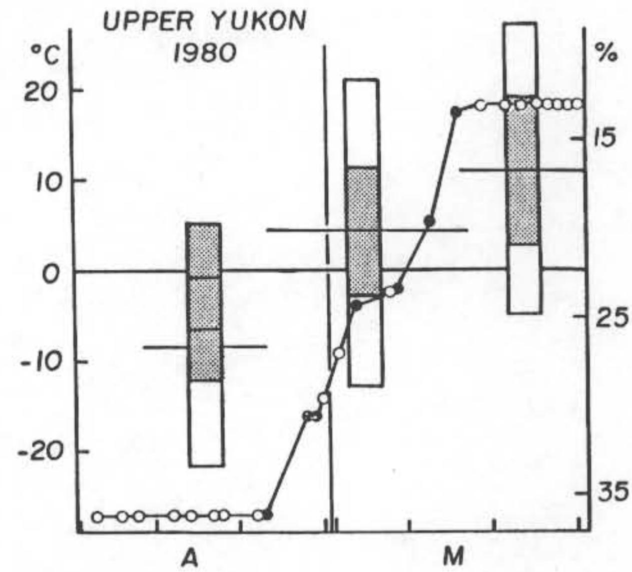

Figure 2. Surface air temperatures and estimated surface albedo in the upper Yukon in the spring of 1980. Albedo measured over the entire region (●) or a representative portion of it (○). Temperatures cover the 14 days prior to albedo decrease, 21 days during melt, and 14 days following melt and include average highs, means, and lows (heavy stippling) and the extreme high and low recorded at stations in the region during the period in question.

The significance of surface albedo as an active factor influencing spring temperatures, and snowmelt in the regions was next investigated. One minus the daily albedo was multiplied by the daily total insolation reaching the top of the atmosphere in a given region and a constant atmospheric screening factor of .6 to obtain Q, an index estimating the amount of short-wave radiation available for surface heating or evaporation on a given day. The screening factor was parameterized from daily solar-radiation data from Barrow and Fairbanks, reported in Holmgren et al. (1973) and Dingman et al. (1980), and daily, top-of-the-atmosphere insolation data.

Qs were studied in relation to daily high temperatures as shown in Figure 3. Results covering all 3 years show the following:

1. In most cases, 3-day averaged maximum temperatures remain below 5°C until Qs reach 260 cal cm^{-2}/day^{-1}.
2. Three-day averaged daily highs are above 5°C in the forested regions once a Q of 260 is reached in all but one case. This is also the situation on the North Slope once Qs reach 340.

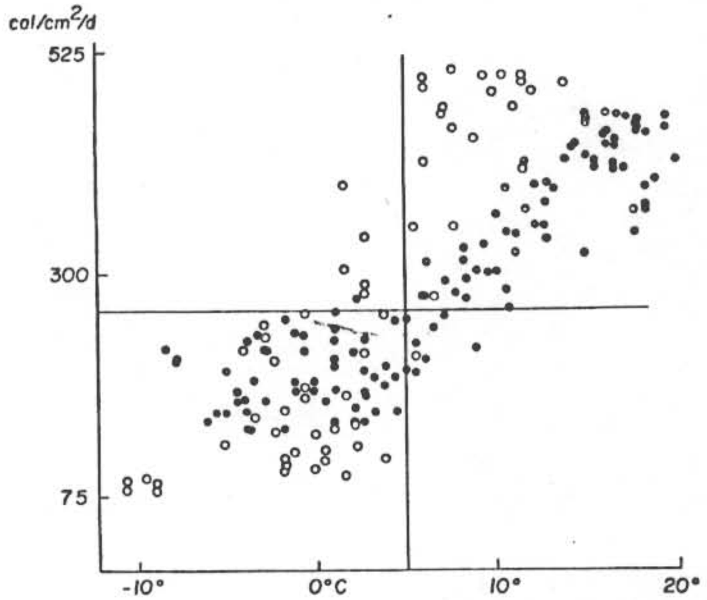

Figure 3. Relationship of temperature to Q in all study regions during the springs of 1979-1981. Each point represents a 3-day average of daily maximum temperatures versus the average Q during the same period. Three-day intervals are between March 20 and May 21 in the two forested regions (●) and April 20 and June 21 in the North Slope region (○).

3. Regional surface albedo does not start decreasing rapidly until Qs reach approximately 260.
4. During the melt interval, 3-day averaged, daily, high temperatures remain between 5 and 13°C in the forested regions and 2 and 10°C in the tundra.
5. Qs are above 415 once snowmelt is completed. At this point, highs remain above 13°C in the forested regions and 5°C on the North Slope.

Spring, maximum, surface-air temperatures in parts of Alaska are, therefore, largely dictated by the attainment of Q thresholds which themselves are functions of insolation and albedo.

Averaging the 3 years, a Q of 260 is reached on the North Slope (~70N) 42 days later than in the Tanana Valley (~64N). This is primarily due to differences in regional albedo, ranging upwards to 45 per cent when both are fully snow covered, and not from the 6° difference in latitude. The 6 degrees amount to a 6-day lag in equivalent daily insolation reaching the top of the atmosphere at 70° compared to 64° in mid-April. This lag gradually diminishes until the latter part of May, when daily insolation at 70° exceeds 64°.

Thresholds may also be reached on different dates within a region from year to year. The Q threshold of 260 is reached in the Tanana Valley on April 21, 1979; April 12, 1980; and April 18, 1981. Figure 4 shows spring temperatures and albedos for this region during these years and suggests the timing differences are the result of various conditioning factors which affect regional albedo prior to threshold attainment. For instance, advection of warm air into the region early in the spring of 1980 began to melt the snow slowly and decrease the albedo, leading to an early threshold data. In the same sense, rain and lack of a heavy winter snowpack can lead to slightly lower albedos and an early date. None of these factors are sufficient to result in major temperature or albedo changes until rising insolation values combine with the slowly falling albedos to reach a Q of 260. In the opposite sense, a heavy snowpack, as was the case in 1979, or a major spring snowfall, as in 1981, will delay the threshold, even counteracting earlier opposite effects, as in 1981. Threshold-timing differences of 13 days were found in the tundra. In the upper Yukon, each spring had a similar deep snowpack which may

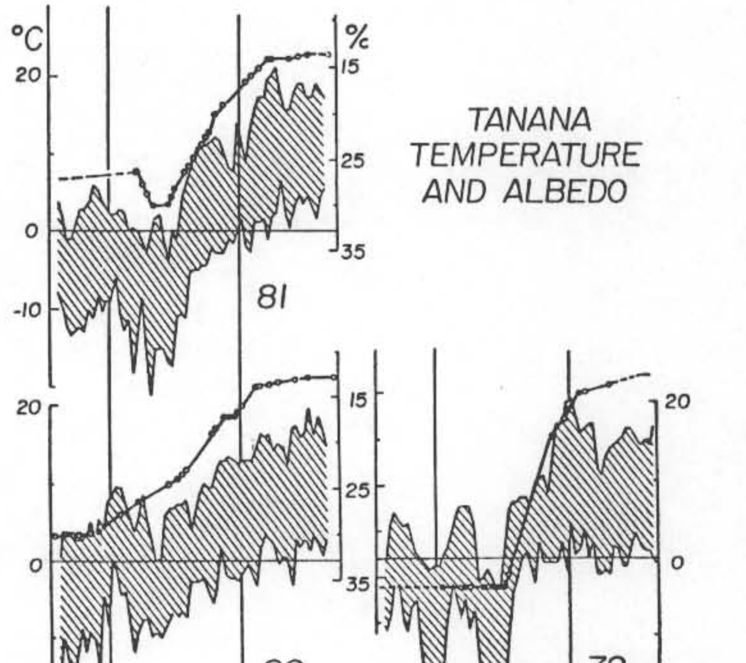

Figure 4. Surface air temperatures and estimated surface albedo in the Tanana Valley in the springs of 1979-1981. Albedo same as in Figure 2. Daily highs and lows averaged from five stations within the region are separated by hatching.

explain why timing differed by only 3 days.

The Q-temperature responses of two local areas on the North Slope were found to be quite sensitive to surface albedo changes. Figure 5 shows the similarity in fluctuations of Q and temperature between the two 50km x 50km blocks in May 1980, prior to the slightly earlier commencement of snowmelt in the Umiat area. The advection of an anomalously warm air mass into the region in late April 1979 caused some melt in both blocks. The resultant albedo decreases differed be-

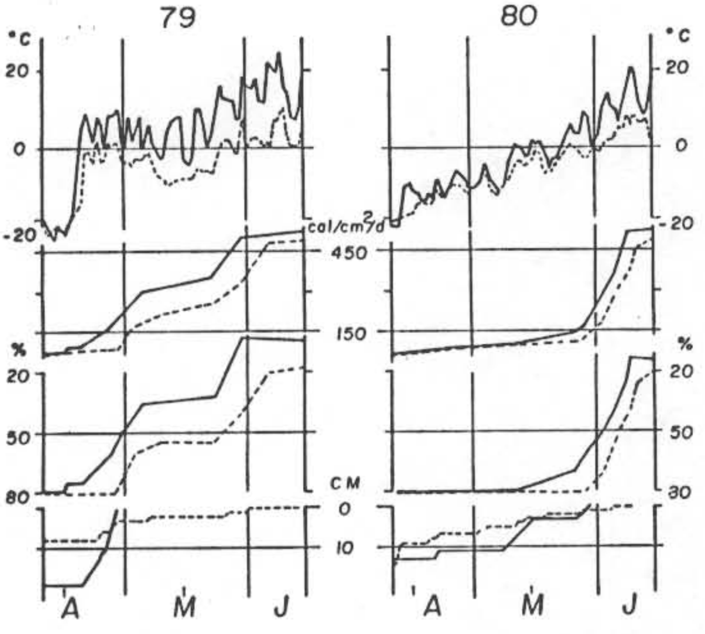

Figure 5. Daily high temperature, Q, estimated surface albedo, and station snow depth (top to bottom) for Umiat (———) and Barrow (- - - -) locales from April 10 to June 15, 1979 and 1980.

tween Barrow and Umiat, leading to higher temperatures at the latter, lower-albedo location throughout May. It is also noted that:

1. The April 1979 melt did not cause albedos to fall to levels where Q thresholds were reached. Therefore, snowmelt continued slowly and temperatures held steady until the latter half of May when thresholds were reached and rates of change increased.
2. Despite the reported loss of snow at the Umiat station on April 29, 1979, significant snow remained in the region 3 weeks longer. The importance of satellite imagery analysis is obvious here (for a thorough description of snowmelt on the Alaskan tundra, see Holmgren et al. 1973).
3. A partial analysis of imagery indicates that the oscillatory nature of the high temperatures, seen particularly at Umiat in May 1979, is due to the presence (cool days) or absence (warm days) of extensive cloud cover. The lack of such a pattern at Barrow is most likely due to a combination of a stronger maritime influence (Dingman et al. 1980) and its higher albedo. The latter reduces cloudy- and clear-day differences in daily, net, shortwave radiation at the surface by increasing multiple reflectons between the surface and cloud base, thus increasing net shortwave radiation on cloudy days (Wendler et al. 1981).

DISCUSSION

While the results of this study cover only 3 years, they provide insight into the potential impact of CO_2 induced change in the spring climate of portions of Alaska.

If the depth of the winter snowpack remains within the range of present-day values and vegetative conditions do not change, a significantly earlier threshold attainment could occur from conditioning factors operating to decrease snow-cover albedo. Factors might include present-day, early-spring snowfalls becoming rain events and, more frequently, advection of warmer air into a region. As an extreme, without snow cover a Q value of 260 would be reached by April 3 in the Tanana Valley and April 11 on the North Slope. With changes probably falling between present levels and these extremes, increases in spring temperature equal to Ramanathan et al.'s (1979) zonally averaged model results of 5 to 6°C could occur. However, an increase in the snowpack as a result of warmer, wetter winters could act to delay or maintain the threshold's present-day timing.

CONCLUSION

Analysis of satellite imagery and ground-station data over portions of Alaska during the springs of 1979-1981 showed a strong asssociation of temperature and estimated surface albedo. Regional albedo, varying as a function of snow cover, was incorporated into an index regionally parameterizing the value of absorbed solar radiation at the surface. Index values of approximately 260 cal cm^{-2} day^{-1} were reached immediately prior to periods of rapid snowmelt and sharply rising temperatures. The timing of this threshold varied between areas as a result of regional surface albedos differing by up to 45 per cent with fully snow-covered ground. Intraregional differences in timing were due to variables such as advected warm air or spring snowfalls acting on the snow cover and conditioning the regional albedo prior to threshold attainment. High temperatures in most cases remained below 5°C prior to reaching the threshold, 5-13°C during rapid snowmelt, and over 13°C following melt. A potential CO_2-induced warming in Alaska could result in significantly earlier threshold attainment if snow depth remains within the range of present-day values, while an increase in cover due to warmer, wetter winters could delay the threshold. o

REFERENCES

Budyko, M. 1969. The effect of solar radiation variations on the climate of the Earth. *Tellus 21*: 611-619.

Dewey, K. 1977. Daily maximum and minimum temperature forecasts and the influence of snow cover. *Mon. Weath. Rev. 105*: 1594-1597.

Dingman, S., R. Barry, G. Weller, L. Benson, E.LeDrew and C. Goodwin. 1980. Climate, snow cover, microclimate, and hydrology. IN: An Arctic Ecosystem: The Coastal Tundra at Barrow, Alaska. J. Brown, etc., eds. Dowden, Hutchinson, and Ross. pp. 30-65.

Holmgren, B., C. Benson, and G. Weller. 1973. A study of the breakup on the Arctic Slope of Alaska by ground, air, and satellite observations. IN: Climate of the Arctic. (G. Weller and S. Bowling eds.) Geophysical Institute, Univ. of Alaska. pp. 358-366.

Kukla, G. 1981. Climatic role of snow covers. IN: Sea Level, Ice, and Climatic Change. Proceedings of the Canberra Symposium, December, 1979. IAHS Publ. 131. pp. 79-107.

Manabe, S., and R. Wetherald. 1975. The effects of doubling the CO_2 concentration on the climate of a general circulation model. *J. Atmos. Sci. 32*: 3-15.

Manabe, S., and R. Wetherald. 1980. On the distribution of climate change resulting from an increase in CO_2 content of the atmosphere. *J. of Atmos. Sci. 37*: 99-118.

McFadden J., and R. Ragotzkie. 1967. Climatological significance of albedo in central Canada. *J. Geophys. Res. 72*: 1135-1143.

Namias, J. 1962. Influences of abnormal surface heat sources and sinks on atmospheric behavior. IN: Proceedings of the International Symposium on Numerical Weather Prediction in Tokyo, 1960. Meteorological Society of Japan. pp. 615-627.

Namias, J. 1980. Some concomitant regional anomalies associated with hemispherically averaged temperature variations. *J. Geophys. Res. 85*: 1585-1590.

NOAA. 1979-81. Climatological Data (Alaska). Environmental Data Service.

Ramanathan, V., M. Lian, and R. Cess. 1979. Increased atmospheric CO_2: Zonal and seasonal estimates of the effect on the radiation energy balance and surface temperature. *J.Geophys. Res. 84*: 4949-4958.

Sellers, W. 1969. A global climatic model based on the energy balance of the earth-atmospheric system. *J. Appl. Met. 8*: 392-400.

Walsh, C., D. Edman, and J. Zapotocny. 1981. GOES observation of a rapidly melting snowband. *Mon. Weath. Rev. 109*: 1353-1356.

Weller, G., S. Cubley, S. Parker, D. Trabant, and C. Benson. 1972. The tundra microclimate during snowmelt at Barrow, Alaska. *Arctic 25*: 291-300.

Wendler, G., F. Eaton, and T. Ohtake. 1981. Multiple reflection effects on irradiance in the presence of Arctic stratus clouds. *J. Geophys. Res. 86*: 2049-2057.