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Department of Geography (Climatology)
McGill University
805 Sherbrooke Street West
Montreal, Quebec, Canada
H3A 2K6

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TOPOGRAPHIC EFFECTS ON WINTERTIME SURFACE PRECIPITATION PATTERNS IN THE GREATER MONTREAL REGION

by

B. Hrebenyk and John E. Lewis*

Introduction

Subjective observations of wintertime precipitation patterns in the greater Montreal region indicate certain anomalies exist as to surface accumulation. One example commonly cited is the tendency for the Eastern Townships area to receive appreciably more precipitation than Montreal during major snowstorms. This pattern is not always the case, yet it occurs with enough frequency to attract attention. Similarly, the north shore of the St. Lawrence River in the vicinity of l'Assomption appears to receive more precipitation relative to the area west of Montreal toward the Ontario border, while residents of Burlington, Vermont, speak of the sheltering effect of the Adirondacks during major storms. No documented evidence exists to support these impressions.

The purpose of this article is to examine surface precipitation patterns from twenty-two major storms for two winter periods (75-76 and 76-77). The patterns of surface accumulations for individual storms and the two season totals are compared to upper level winds and storm characteristics. A qualitative assessment is made of the degree to which precipitation patterns depend on the direction of airflow relative to major topographic features in the study area. A later article, as a complement to surface

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* B. Hrebenyck is a graduate of the Department of Geography at McGill University currently employed as a climatological consultant in the siting of power plants. John E. Lewis is Associate Professor in the Department of Geography at McGill University.

accumulation patterns, will examine the frequency occurrence of precipitation echoes computed from radar imagery to ascertain whether preferred regions for synoptic precipitation generation within the area exists.

Study Region

The 500 m contour line was chosen as a reference level because it best describes the nature of the two major mountain ranges to the south and southeast of Montreal. The Adirondacks of New York State form a fairly massive barrier to airflow with maximum relief of approximately 1500 m. However, most of the range lies between 500 m and 900 m elevation. Maximum elevations in the Green Mountains of Vermont are only slightly lower (≃ 200 m), but the range itself forms a distinct series of narrow parallel ridges which are clearly depicted by the 500 m contour. The extent of the study area can be seen on Fig. 1* and areas above 500 m are blackened.

The Laurentian Shield formation which lies to the north of Montreal is less of a mountain range than an extensive plateau of undulating terrain whose edge lies parallel to the north shore of the Ottawa River in the west and the St. Lawrence River in the northeast. The 500 m contour in this case does not properly describe this formation. Most of the shield area lies between 200 m and 500 m, and the 200 m contour (not shown on Fig. 1) along the edge of the river valleys better defines the limits of this formation.

To the west-southwest of Montreal lies a broad region of lowlands stretching all the way to Lake Ontario. The convergence of the Ottawa and St. Lawrence rivers properly outlines the narrowing of this region into the St. Lawrence valley at the junction of the two rivers. If the 200 m contour is assumed to represent the edge of the valley systems, then the average width of the St. Lawrence valley would be about 100-120 km, being its narrowest at 80 km roughly along 74°W (longitude). Adjoining the St. Lawrence valley is the narrow valley between the two southern mountain ranges which contain Lake Champlain. Average width of the valley is 25-30 km.

Relevant Literature

Studies by Colucci (1976) and Tyner (1970) have indicated that major topographic features of eastern North America materially affect the development and movement of cyclonic storms. Browning, Pardoe and Hill (1975), Bader and Roach (1977), Wilson and Atwater (1972), Clodman and Jarvis (1966),

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* For convenient reference, all figures have been placed at the end of this article.

Pittock (1977), and Myers (1964) all attest to the significance of interactions between large-scale circulation and local topography in explaining meso-scale precipitation variability. Harrold and Austin (1974) describe the dynamical controls of precipitation systems which are largely controlled by the motion of the air producing the precipitation; therefore precipitation systems should be classified according to the types of flow which produce them. Hrebenyk (1980) reviews very succinctly the general literature on topographic effects and meso-scale organization of precipitation patterns.

Only one study of the spatial distribution of winter-time precipitation has been attempted for the greater Montreal region. Carlson (1968) showed that, for dry snow at least, it was possible to produce reasonable estimates of surface accumulations from radar derived snowfall rates. However, Carlson's purpose was to determine the accuracy of radar estimates and did not address the topographic effects per se related to the synoptic character of storm types.

Bellon and Austin (1977) have described topographic influence on the development and decay of summertime storms. Three areas of distinct upslope flow are indicated in their study:

1) Windward sides of the Laurentians (i.e., that area defined by the 500 m contour on Fig. 1;
2) the Adirondacks; and
3) the Sutton Hills of the Green Mountain range.

Decay of storms was prevalent along a narrow corridor on the eastern edge of the Laurentians, which represents a lee wave effect of the type described by Clodman and Jarvis (1966). Lakes Champlain and St. Pierre appeared to dampen storm development also. Powe (1968) states that a secondary maximum of wind direction frequency occurs from the north-northeast for Dorval which is produced by the St. Lawrence Valley channelling surface winds. During periods of precipitation, the colder temperatures over the valley resulting from this flow cause snowfall to persist a little longer here than in surrounding regions. St. Hubert, a few miles east of Montreal, has slightly greater frequency of winds from the south, which according to Powe, displays a weak effect of valley wind along the Richelieu River.

Allard (1974) simulated northwesterly wind conditions in the region to examine topographically induced flow phenomena. He found that over the Laurentian plateau, vertical motion remained fairly small, while the northern edge of the St. Lawrence valley exhibited a slow sinking motion in the airflow and a rising motion was evident along the southern edge. A generally southwesterly airflow predominated for the storms we studied during the two winters; therefore the pattern of vertical motion described by Allard would be reoriented with respect to the change in flow direction.

A contrasting view as to precipitation controls stated in the preceding references is that of Carlson (1968) and Drufuca (1977) who contend the "geographical" effects on precipitation in the Montreal region are minimal.

Storm Periods and Precipitation Data

Leduc (1977) concluded that for Montreal the seasonal total precipitation from November through April is, to a large extent, determined by the number of major snowstorms that occur. In other words, the more frequent occurrences of light precipitation actually account for very little of the total precipitation in the region. Heavy snowstorms ($\geq$ 1.52 cm water equivalent) accounted for about 70% of the total snowfall, and rainfall over 5.08 mm accounted for more than 80% of the total rainfall during the winter.

Applying a similar criteria, it was decided to limit the study to only the most intense storms ($\geq$ 0.5 ins precipitation) which simplified the analysis but accounted for a major portion of total precipitation. Table One shows the 22 storms investigated. Thirteen of the storms were the result of the passage of single cyclone centers. The remaining nine storm periods consisted of two or more cyclone centers resulting in protracted periods of precipitation which could not be differentiated into separate storms. In all, thirty-four separate cyclonic centers were involved with only a slightly greater tendency for the storm centers to be located along the east coast. The precipitation data employed to compile the resulting maps was extracted from published summaries of the Monthly Record for the Canadian stations and from the Climatological Data supplied by NOAA for New York and Vermont. A total of 118 stations was used.

It is recognized that due to the inherent difficulty in measuring wintertime precipitation, errors are inevitable. Since the nature of the study is to compare general patterns rather than actual quantitative amounts, the effects of the errors in measurement will be minimized so long as comparisons are restricted to relative amounts (i.e., fairly broad categories). No distinction is made between rainfall and snowfall. Hrebenyk (1980) conducted a careful analysis as to possible station errors and representativeness of the interpolated isohyetal maps of surface accumulations.

Patterns of Surface Accumulations

The surface accumulation patterns were analyzed and then grouped

TABLE ONE

Occurrence of Storms Studied in the Montreal Area

Storm Period	Duration	No. of Cyclones Active During Storm Period
November 7-8, 1975	24	1
November 10, 1975	15	1
November 12-14, 1975	50	3
December 9-10, 1975	25	2
December 17-18, 1975	27	2
December 25-26, 1975	17	1
December 30, 1975	18	1
January 13-14, 1976	15	1
January 25-27, 1976	52	3
February 1-2, 1976	16	1
February 18-19, 1976	26	2
February 21-22, 1976	20	1
March 21, 1976	20	1
March 31- April 1, 1976	40	3
December 6-7, 1976	16	2
January 10, 1977	16	2
February 24-25, 1977	21	2
March 4-5, 1977	15	1
March 13-14, 1977	36	1
March 22-23, 1977	23	1
April 2-3, 1977	20	1
April 4-5, 1977	27	1
TOTAL	539 hours	34

according to the similarities of the patterns. In this manner, five groupings were identified, and the patterns in three of these were found to be related to the airflow at the 700 mb and 850 mb levels while the fourth and fifth groups were composed of specific synoptic situations. Quite naturally, the more complex storms in which winds at upper levels varied greatly tended to produce patterns which were unique and which could not be included in such groupings.

In the following sections, one storm from each of the five groups is presented to illustrate the types of patterns observed. It is understood that some of the spatial differences in precipitation amounts may not be significant and may have arisen from inaccurate measurements of actual precipitation. The meanderings of some of the isohyets is clearly open to question. It is in the overall location of areas with precipitation amounts above the median value (depicted by stipling) which is the main concern because it is the similarity in the location of these areas for different storms which defines each group.

Case of January 10, 1977 (Group 1)

This storm consisted of two low pressure centers, one of which moved toward the Great Lakes to the west of the Appalachians while a secondary low moved up the Atlantic seaboard of the United States. In the early part of the storm, the western center predominated while in the latter half the East Coast Low acted as the focus of the storm. Winds at 850 mb. shifted slightly from 155° at the start of precipitation to 135° at the end, while 700 mb. level winds remained steady from the south at 190° throughout the storm.

Fig. 1, showing the surface accumulation pattern, indicates that the bulk of the precipitation was deposited north and west of the St. Lawrence River with the heaviest falls occurring to the northwest of Montreal. Large amounts of precipitation were also deposited on the windward side of the Adirondacks to the 850 mb. winds, while on the lee side of the range, between the mountains and the St. Lawrence River, a distinct shadow effect is evident. A similar situation exists on the windward and lee slopes of the Green Mountains as well, with Montpelier receiving over 1.0 inches (25.4 mm.*) of precipitation compared with only 0.5 inches (12.7 mm.) at Burlington.

A storm featuring almost identical synoptic conditions for December 9-10, 1975 produced essentially the same pattern of deposition suggesting that the deposition in both storms was not entirely random, i.e. not determined entirely by the random influx of moisture and instability. The same windward/leeward effects were noted in the December storm for the Adirondacks and Green Mountains. More significantly, however, the deposition of the bulk of the precipitation in the December storm also occurred over the Laurentians and in the region between the St. Lawrence and Ottawa rivers near Ottawa while the region lying between the St. Lawrence River and the Adirondacks and Green Mountains had generally lower precipitation. This same basic pattern was repeated for a third storm on March 22-23, 1977 for an East Coast Low. In all three cases, winds at 700 mb. were southerly while at 850 mb. they were southeasterly.

Case of February 24-25, 1977 (Group 2)

Fig. 2 shows the pattern of precipitation at the surface which

* All published data on total precipitation was obtained in British units, and analysis was preformed on the basis of these units. All results are presented in British units with their metric equivalents in brackets.

resulted from a low center that developed in the mid-west and tracked across Lake Superior toward James Bay. A westerly track was common for all storms in this group. Mainly however, it is the direction of airflow at 700 mb. and 850 mb. which best characterizes the surface patterns in the group.

For the storm of February 24-25, 1977, winds at 700 mb. ranged from 190° at the start of the storm to 165° at the end while at 850 mb. winds were almost constant at 180°-185°. Thus, the main difference between this storm and those in the previous group described above is while 700 mb. winds remain generally southerly, 850 mb. winds are southerly as opposed to southeasterly. In this case, the bulk of the precipitation was deposited over the Laurentians directly north of Montreal, while the western portion of the study area toward Ottawa was generally excluded from the zone of highest precipitation totals. Once again, however, there is an obviously lower amount deposited in the region of the St. Lawrence valley between the St. Lawrence River and the two southerly mountain ranges. The windward/leeward effects within the Lake Champlain basin are not evident, but this is understandable since the airflow at 850 mb. was along the axis of the valley. The pattern which does appear in this region showing the southern half of the basin having received substantially greater amounts of precipitation than the northern half of the basis was consistent for all cases of southerly airflow at 850 mb.

Three other storms produced similar patterns of precipitation to that of Fig. 2, though in each case there was a southwesterly component to the 700 and 850 mb. winds toward the end of the storms which was reflected in a slight increase in deposition in the region to the east of Montreal. Nevertheless, the main points held true in that in the shift from southeasterly flow at 850 mb. in the previous group to the more southerly flow represented by Fig. 2, the main area of highest precipitation deposition shifts from being more or less beyond the St. Lawrence River to the west and north of Montreal to a region more directly to the north. In each case there remains an area of lower precipitation deposition between this region of high depositions and the southerly mountains.

Case of March 21, 1976 (Group 3)

Fig. 3 typifies the pattern of deposition characteristic of storms in which 850 mb. winds start out southwesterly in the early part of the storm and gradually shift to a more westerly flow during the storm while 700 mb. winds remain southwesterly throughout the storm. Precipitation in this storm resulted entirely from the passage of a cold front from west to east across the study area. Winds at 850 mb. shifted from 236° to 287° between the beginning and end of precipitation, and the more intense part of the storm occurred later in the storm when 850 mb. winds were more westerly. The westerly component in wind direction was very prominent in each of the other two cases identified in this group. However, whereas the storm of March 21, 1976 involved a westerly low center, the storms of December 17-18, 1975 and December 6-7, 1976 began with a westerly low, but switched to east coast lows for the latter parts of the storms. Clearly the surface synoptic conditions which produced the similar patterns of surface deposition for these three storms are not consistent. The common factor in all three is the direction of upper level winds.

The major difference between the pattern in Fig. 3 and that of Figs. 1 and 2 is that the major deposition of precipitation occurs south and east of the St. Lawrence River, diametrically opposite to the pattern for southeasterly flow at 850 mb. In all three cases in this group, two distinct regions of high deposition were observed: 1) on the western (windward) side of the Adirondacks, and 2) over the eastern half of the St. Lawrence valley. The two regions were consistently separated by a narrow zone of lower surface accumulations which was an extension of a much larger area of low surface accumulations in the lee of the Adirondacks. It is thought that the higher accumulations over the eastern half of the St. Lawrence valley in Quebec reflect a further shift in the locus of high deposition which was mentioned in a previous case (February 24-25, 1977) as arising from a southwesterly component to the 850 mb. winds. On the other hand, the high deposition on the windward slopes of the Adirondacks is believed to result from the westerly component of the upper level winds because it was not found to be present unless westerly winds occurred. Windward slopes of the Green Mountains probably are affected as well to some extent, but it does not exhibit the same marked increase in precipitation quite possibly because it lies downwind of the Adirondacks for purely westerly flow.

Case of December 30, 1975 (Group 4)

The fourth group of storms, represented by Fig. 4 comprises the set of storms which tracked directly up the St. Lawrence valley and is most analogous to the storms investigated by Carlson (1968). These storms were found to result in patterns of deposition which were in some ways distinct from all others.

In the case of December 30, 1975, the storm tracked directly over Ottawa in a line just south of the Laurentians following the north shore of the St. Lawrence River out of the study area. Winds throughout the storm were southwesterly at upper levels ranging from 256° to 233° at 700 mb. and 245° to 229° at 850 mb. Highest precipitation deposition coincided with the track of the surface low center. The only other notable feature of the deposition was the extensive shadow cast by the Adirondacks over the whole of the Lake Champlain basin.

The prevalence for isohyets to lie roughly parallel to the St. Lawrence River was noted by Carlson (1968) and was a feature of all three storms investigated in this study. In the case of the other two storms in this group, January 13-14, 1976 and February 21-22, 1976, the storms took a slightly more southerly track with the low centers passing directly over the Adirondacks and Lake Champlain. Both storms were accompanied by southeasterly winds at 850 mb., as opposed to southwesterly winds for the case of December 30, 1975, and in both cases the highest surface depositions parallelled the storm track but occurred to the left of the track. In this respect, the two storms were in close agreement with the results reported by Gutzman (1972) that for heavy snowfalls in eastern Canada the areas of heaviest precipitation are generally elongated in the direction of the storm track and occur within 2° latitude to the left of the track of the surface low. Based on the three storms examined in this study, Gutzman's findings may be qualified by stating that the displacement to the left of the storm track will not occur if upper level winds blow nearly parallel to the direction of movement of the storm. However, Gutzman's pattern will occur if the direction of airflow at 850 mb. is at some angle to the surface storm track.

The southeasterly component of the 850 mb. flow in the January and February storms resulted in the characteristic windward/leeward effects for the Adirondacks and Green Mountains described in Fig. 1 within the Lake Champlain basin. Thus, for all storms in this group, at least part of the Lake Champlain basin experienced lower precipitation totals. The effectiveness of the Green Mountains in creating a precipitation shadow, however, was not as pronounced for the southeasterly winds as was the effect of the Adirondacks for the southwesterly winds depicted in Fig. 4.

Case of February 1-2, 1976 (Group 5)

Two storms in the study were typical of East Coast storms in which there is a fairly sharp boundary between higher precipitation in the eastern portion of the study area and lower precipitation in the western portion of the area. Of the two, December 25-26, 1975 and February 1-2, 1976, the latter was the more dramatic. The precipitation in this storm resulted from a rapidly moving east coast low which developed in the Gulf of Mexico and within 36 hours had reached New Brunswick. Winds at 700 mb. remained generally southerly at between 212° at the start of the precipitation and 188° at the end. Winds at 850 mb. shifted enormously during the storm from 207° at the start to 322° at the end.

Fig. 5 depicts the surface accumulation in the study area. It is quite obvious that moisture advection during the storm was limited to the eastern portion of the study area. There is such a sharp gradient between high and low precipitation that it is easy to imagine how in such cases accurate prediction of precipitation amounts along the boundary would be quite impossible. Furthermore, while the highest precipitation amounts clearly lie in the east, the various lobes of low precipitation amounts near Lake Champlain, in the Eastern Townships, and along the St. Lawrence River northeast of Montreal present a fairly confused pattern which makes the exact delineation between high and low precipitation difficult to place. Given the manner in which storms were chosen for inclusion in the study (i.e. a minimum of 0.5 inches of precipitation at Montreal), it is quite likely that a number of such storms were left out of the study because of insufficient precipitation at Montreal. In the case of February 1-2, 1976, the 0.5 inch isohyet passes right through Montreal. In the other case, December 25-26, 1975, the isohyet is only 50 km. from the McGill University station to the northwest.

In the case of Fig. 5, none of the typical windward/leeward effects noted in other storms was evident, perhaps because of the wide shift in the direction of flow at 850 mb. However, in the December storm, which was characterized by east-southeasterly winds, such effects were clearly evident within the Lake Champlain basin. Nevertheless, in both cases the overriding element which determined the patterns of deposition was the sharp boundary between higher moisture deposition in the eastern half of the study area and low moisture in the western half.

Non-conformal Storms

The five groups of storms discussed above accounted for 15 of the 22 storms investigated. Of the remaining seven, five cases were sufficiently complex, either with respect to the number of low centers active during the period of precipitation or with respect to shifts in upper level winds, to be considered unique storms. The remaining two storms had upper level and surface synoptic conditions similar to those represented in either Figs. 1 or 3, but failed to produce surface accumulations which were entirely consistent with either of these two groups. Only some parts of the patterns were reproduced, particularly with regards to windward/leeward slopes of the mountain ranges. However, significant differences in the overall patterns made these storms unlike any within the two groups. These two storms must be regarded as the exceptions to the rule, and as representing the unpredictable elements of some storms.

Correlation of Precipitation Patterns with 850 mb. Airflow

Although the direction of airflow at both 700 mb. and 850 mb. was important in the establishment of the groupings represented by Figs. 1-3, it is evident that the best indicator of the pattern of distribution in these three groups was in fact the flow at 850 mb. For example, the difference between Fig. 1 and Fig. 2 was mainly the fact that in the former case the 850 mb. flow was from the southeast while in the latter case it was from the south. In both cases the 700 mb. flow was essentially southerly. Furthermore, all cases of anomalous windward/leeward distributions around mountain ranges were associated with the flow at 850 mb. The western slopes of the Adirondacks were not observed to display higher precipitation deposition unless there was southwesterly flow at 850 mb. Similarly, the eastern slopes of the Adirondacks received higher precipitation for southeasterly flow. No such effects were observed to occur for southerly flow indicating that in the former two situations the Adirondacks obstruct the flow while in the latter case the flow is along the sides of the range.

Table Two demonstrates the extent of the windward/leeward differences associated with the airflow at 850 mb. The storms refer to groups one and three for SE and SW 850 mb. airflow respectively as well as the two nonconformal storms which were distinct exceptions. Clearly the differences for the Adirondacks are too great and too consistent to be entirely accidental. A similar comparison for the Green Mountains is not as successful, however. For southeasterly flow, a distinct shadow effect is discernable within the Lake Champlain basin when compared to the precipitation within the mountains but for southwesterly flow there is in fact no leeward side to the Green Mountain range in the study area. The narrowness of the ridges and valleys within the range and the scale on which the precipitation data is assembled

TABLE TWO

	ADIRONDACKS		GREEN MOUNTAINS	
	WINDWARD	LEEWARD	WINDWARD	LEEWARD
SE FLOW	Elizabethtown	Lawrenceville	Montpelier	Huntington Center
Storms represented by Fig. 1				
Jan. 10, 1977	1.25 (31.75)*	0.48 (12.19)	1.07 (21.18)	0.49 (12.45)
Dec. 9-10, 1975	0.62 (15.75)	0.35 (8.89)	0.37 (9.40)	0.17 (4.32)
Mar. 22-23, 1977	1.04 (26.42)	0.54 (13.72)	0.45 (11.43)	0.25 (6.35)
Non-conformal storm				
Dec. 25-26, 1975	1.20 (30.48)	0.70 (17.78)	1.12 (28.45)	0.28 (7.11)
SW FLOW	Colton	Plattsburg	Bristol	Montpelier
Storms represented by Fig. 3				
Mar. 21, 1976	0.73 (18.54)	0.38 (9.65)	0.40 (10.16)	0.42 (10.67)
Dec. 17-18, 1975	0.36 (9.14)	0.10 (2.54)	0.21 (5.33)	0.03 (0.76)
Dec. 6-7, 1976	1.34 (34.04)	0.40 (10.16)	M (1.30 E) (33.02)**	1.19 (30.23)
Non-conformal storm				
Nov. 7-8, 1975	0.84 (21.34)	0.20 (5.08)	0.17 (4.32)	0.71 (18.03)

* in inches (mm.) of total precipitation

** based on 1.19 inches (30.23 mm.) at Huntington Center and 1.52 inches (38.61 mm.) at Cornwall.

cannot accurately reflect the finer structure of precipitation within the mountains. In addition, the effectiveness of the Adirondacks in creating a precipitation shadow within the Lake Champlain basin interferes with the windward enhancement of precipitation on the western slopes of the Green Mountains. Consequently, the windward/leeward effects for these mountains are seldom observed for southwesterly flow situations.

The extent of the precipitation shadow cast by the Adirondacks was estimated at roughly 40-80 km. in most cases. In the example of Fig. 4, the shadow was somewhat larger in the order of 80-100 km., and was the largest such feature observed. The Green Mountains, by comparison, were observed to cast a shadow in their immediate vicinity within the Lake Champlain basin more typically of about 50 km. or less in extent.

Airflow at 850 mb. was also closely related to the distribution of high precipitation values within the St. Lawrence valley. The ten storms represented by Figs. 1-3 show a consistent shift of the major surface accumulations within the valley as the airflow at 850 mb. changes from southeasterly to southerly to southwesterly, irrespective of the location of the surface low center. For southeasterly flow, heavy precipitation tends to occur in a broad band north and west of the St. Lawrence River. On the other hand, for southerly flow, the zone of major accumulations occurs north and northeast of Montreal while significantly less precipitation occurs west of Montreal. In both cases, the areas of heavy precipitation are separated from the mountainous regions of the Adirondacks and Green Mountains by a distinct zone of lower precipitation amounts typically 40-80 kms. wide. For storms featuring southwesterly flow, the zone of heavy precipitation in the valley is shifted further eastward and lies more or less downwind of the Adirondacks and is characteristically separated from these mountains by what has been described as the shadow effect of the Adirondacks. In all three cases, the zone of heavy precipitation appears to be located downwind of mountainous regions

Fig. 6 was constructed in order to determine whether any of the individual storm features could be discerned in the precipitation patterns resulting from the precipitation pattern derived from the 22 storm periods during the two winters. It can be seen that the variability of spatial precipitation deposition is greatly diminished on the overall maps as compared with individual storm maps and is restricted mainly to two areas. A small region in the Laurentians and an area in the Eastern Townships. The former region has significantly higher precipitation amounts than the surrounding stations. Similar above average amounts in this region are observed in individual storms (Fig. 2) in which prevailing 850 mb. winds were southerly. Therefore, it is possible that the above average values in this region may result from a slight windward enhancement of precipitation by the Laurentians, but the influence is not nearly as obvious as for the Adirondacks. The prominent region of high precipitation on Fig. 6 which occurs in the Eastern Townships is centered around Granby. It is, indeed, noticeable that in this region the steep gradient between low values (<30.0 inches, or <76.2 cm.) within the Lake Champlain basin and high values (>40.0 inches, or >101.2 cm.) in the Eastern Townships, reminiscent of patterns resulting from predominantly southwesterly flow at 850 mb. (Figs. 3), remains as a distinctive feature. Since the prevailing airflow in this region is undoubtedly southwesterly over prolonged periods of time, the persistence of this feature must necessarily be the product of systematic topographic control of precipitation mechanisms.

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This sequence of figures shows precipitation at the surface on different dates. Isohyets for Figs. 1-5 are in units of 0.1 in. (2.54 mm.) and for Fig. 6 in units of 1.0 in. (25.4 mm.). Stippling denotes areas with precipitation above median value.

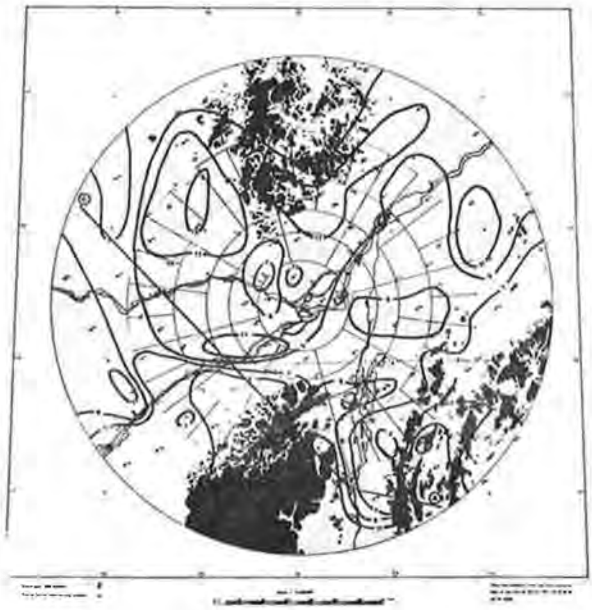

Fig. 1. January 10, 1977. Wind southeasterly.

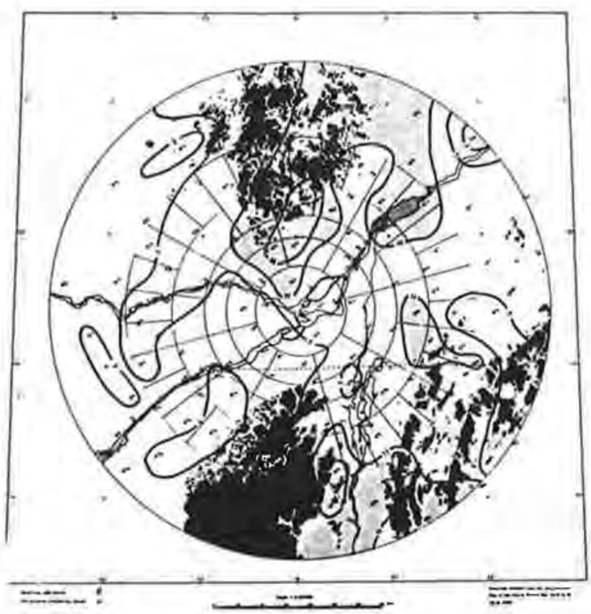

Fig. 2. February 24-25, 1977. Wind southerly.

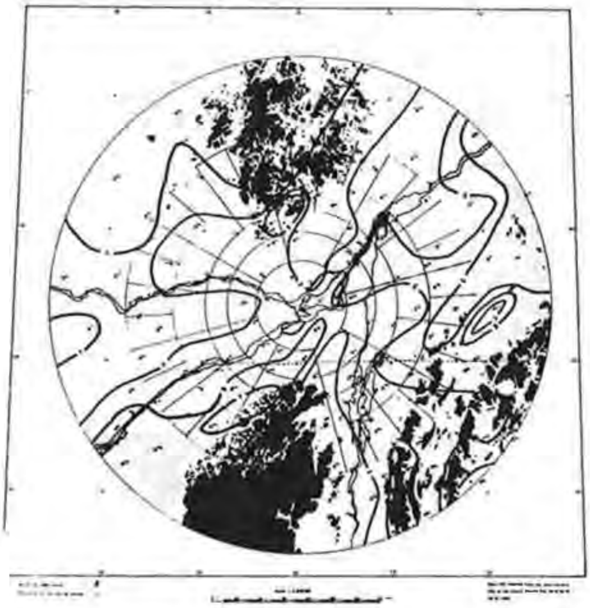

Fig. 3. March 21, 1976. Wind southwesterly.

Fig. 4. December 30, 1975.

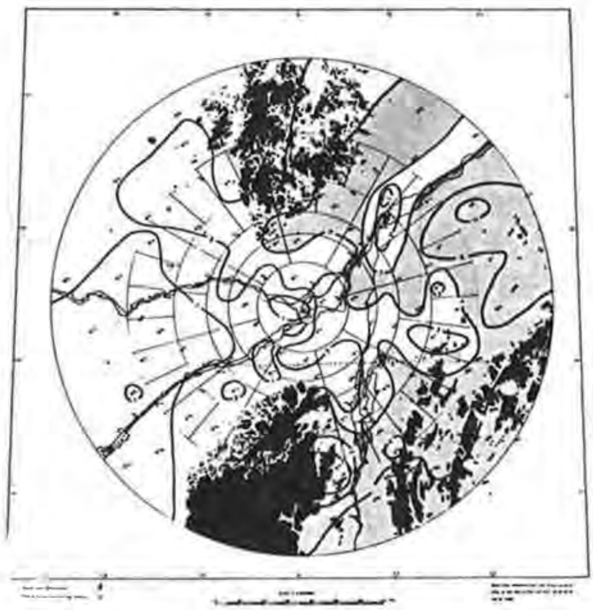

Fig. 5. February 1-2, 1976.

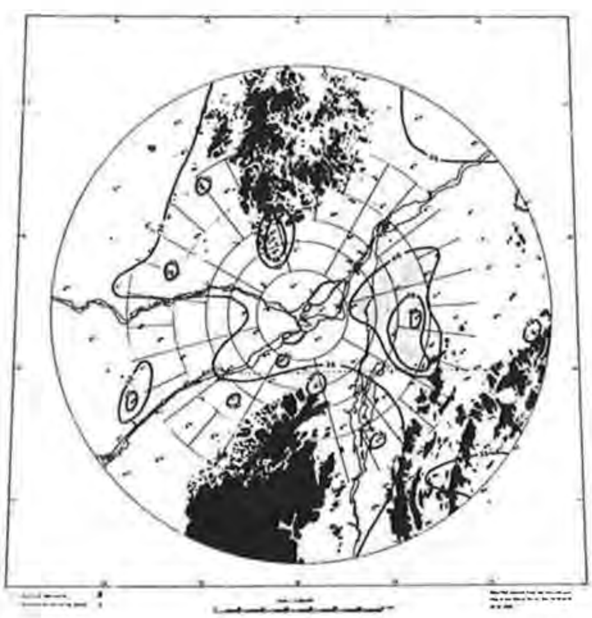

Fig. 6. Distribution of Total Surface Precipitation, November, 197-- to April, 197-- inclusive.

NB. Isohyets in units of 1.0 in. (25.4 mm.).

COMMENTS ON A POSSIBLE RELATIONSHIP BETWEEN DIFFERENTIAL RADIATION BALANCES AND RAINFALL PATTERNS IN BARBADOS

by

Peter Wilson*

Introduction

There often occurs on the west side of the island of Barbados a pattern of rainfall characterized by parallel east to west lines of rain showers separated by clear sky. Three or four such rain 'bands' may occur simultaneously over a ten mile section of the coast, usually at night.

Evidence exists to suggest that the convective instability necessary to produce this rainfall pattern is influenced by the existence of different radiation balances between the generally steep, east-facing, sparsely vegetated slopes of the east coast of Barbados and the gentle, vegetation-covered, west-facing slopes of the western half of the island.

Physiography

Barbados is twenty-one miles long from north to south and fourteen miles across at its widest point. Its six physiographic regions are illustrated in figure 1.

From the highest point in the central uplands (Mt. Hillaby, 1115 ft.) the land descends north, south and west in a series of gently-sloping terraces. The Vale, lowland terraces, eastern lowlands and Christ Church

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* Peter Wilson is a graduate student in the Department of Geography at McGill University.

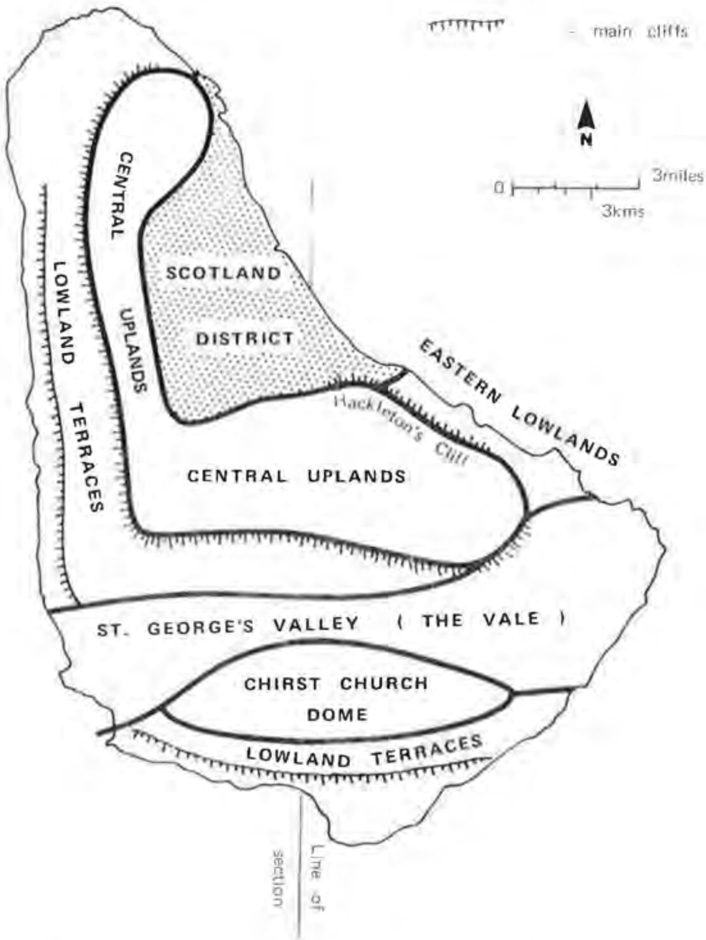


Fig. 1. Physiographic Regions of Barbados (after Macpherson, 1974).

TABLE ONE

Classification of Slopes in Barbados

Slope	% Total Area	Percentage facing quadrant centred on: East	West	South	North
<5°	68	11	---- 70 -----		19
5°-10°	31	25	27	23	25
>10°	1	46	---- 24 -----		30

Dome each represent distinct physiographic regions but are all characterized by gentle slopes. In fact, aside from the amphitheatre-shaped Scotland district and its immediate surroundings on the east coast, Barbados is a flat island, especially when compared with its volcanic neighbours. Only one percent of the total land area of Barbados slopes more than ten degrees, while almost seventy percent of the land slopes five degrees or less. Table One (Garnier, 1975) summarizes the extent and orientation of the slope of Barbados' terrain.

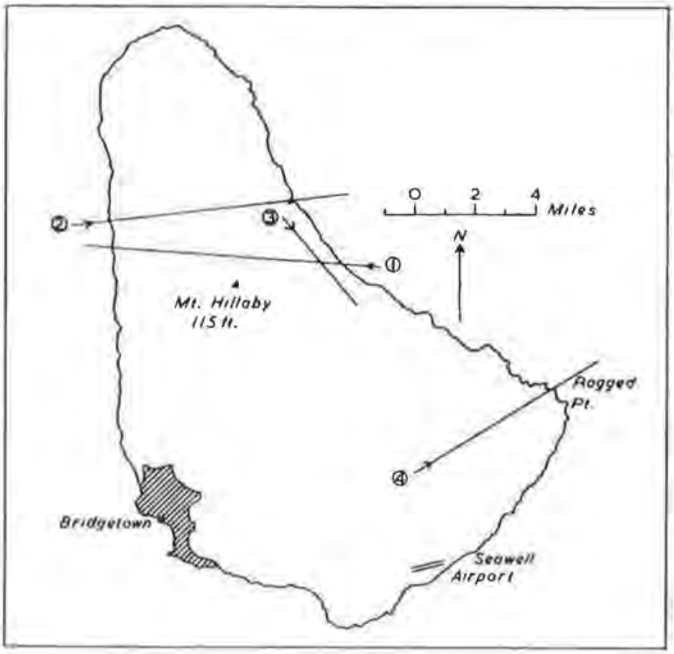
It is the differential diurnal heating of the predominantly east-facing slopes of the Scotland district and its surroundings, as compared with the more gentle, westward-sloping terraces of the west coast that is thought to influence the rainfall pattern described in the introduction. The physiography of the Scotland district will, therefore, be discussed in greater detail in a later section.

The Radiation Balance of the East-Facing Slopes of Barbados' East Coast

Table One illustrates that, as a rule, gentle slopes in Barbados face west while steep slopes are predominantly east-facing. Furthermore, the great majority of east-facing slopes of the island are concentrated in the Scotland district and the area immediately surrounding it. In fact, 37% of that area has slopes with azimuths between 045° and 135°. This is of particular climatological significance in that the azimuths of these slopes embrace the relatively slight seasonal variations in the arrival direction of both the trade winds and the morning sun. Together these slopes make up 1.2 square miles of land, with gradients as follows (Garnier, 1975): less than 11°, 13% of the area; 11°-16°, 68% of the area; 17°-22°, 18% of the area; and more than 26°, 1% of the area.

In the summer of 1973 Garnier (op.cit., 1975) conducted a study to determine to what degree and at what times of the day the surface radiation balance of the east-facing slopes of Barbados' east coast differed from conditions elsewhere on the island. Fourteen sites were established to measure surface radiation. Ten were at Cambridge, in the east coast study area and the other four were at the Caribbean Meteorological Institute on the island's west side.

Garnier's ground observations were supplemented by a programme of remote sensing that employed a Barnes precision radiation thermometer (PRT-5) attached to the step of a Cessna 150 aircraft. The aim of this aerial work was to obtain information on surface variations in radiative



1. East-west; 2. West-east; 3. "North-south"; 4. West to Ragged Point

Fig. 2. Remote Sensing Traverse Lines (after Garnier, 1975).

temperatures along the flight lines illustrated in figure 2.

It was the temperature profiles obtained using the airborne PRT-5 that proved to be the most interesting.

From those profiles Garnier drew the conclusions given below:

1. During the period immediately before sunrise there is little contrast between surface radiative temperatures on the east and west sides of Barbados.

2. The early morning sun, or even the sky diffuse radiation, readily warms the eastern side so that by 0700 hours the surface of east coast area is one or two degrees centigrade warmer than the west.

3. The greatest contrast between east and west surface radiative temperatures occurs between 0700 and 0830 hours. During this period eastern Barbados emperiences rapid heating compared to the western part of the island.

4. The sequence of events described above does not occur in reverse during the twilight hours. The setting sun does not heat the west coast more than the east coast. A PRT-5 profile mirroring that of the morning hours is thus not established. This is attributed to different surface characteristics between the east and west coasts, as well

as to the lower transmissivity of the atmosphere in the afternoon. The result is that by 1830 hours the profiles resemble those of pre-dawn.

5. A "hot spot" was observed to occur shortly after 0700 hours in an area of the east coast region that slopes toward the northeast with a gradient in excess of ten degrees. This combination of slope and azimuth, as well as the low albedo (0.13) of its sparsely vegetated, stoney surface, seems to have produced favourable conditions for rapid heating after daybreak. On at least one occasion observed by Garnier this "hot spot" was seen to influence cloud development. It is thought that under calm synoptic conditions, the consequence of this influence could last throughout the day.

6. Shortwave radiation income was seen to differ between east-facing and west-facing slopes <u>within</u> the east coast study area. Since both the contributions of albedo and sky-diffuse radiation for the two slope orientations can be considered equal, only contrasts in direct solar radiation were recorded. The data indicate that the heating of the east-facing slopes exceeds that of the west-facing slopes until shortly before noon, with a maximum difference between about 0800 and 1000 hours. From 1230 hours onwards the heating of the west-facing slopes exceeds that of the east-facing slopes.

7. The only way in which the surface heating of the east coast area could influence the development of rainfall patterns over the western half of Barbados is by encouraging or discouraging convective instability under relatively undisturbed synoptic conditions.

In order to define the term "undisturbed synoptic conditions" the present article has adopted the precipitation mode system of rainfall pattern analysis used by LaSeur (1965). "Undisturbed synoptic conditions" are considered to exist in Barbados when 55% or less of the island's precipitation reporting stations receive rain in a twenty-four hour period. This occurs 43% of the time during the wet season (but provides only 8% of the rain) and 65% of the time during the dry season (Garstang, 1966).

Convective Instability

Conventional orographic and heat island models fail to explain the rain bands over Barbados. Both of these models predict heavy <u>daytime</u> precipitation in the lee of either physical or convective elevators of moist air. The challenge confronting climatologists, therefore, is to explain what Garstang (1966) refers to as "...the dominance of a nighttime (rainfall) regime and the absence or weakness of any daytime maximum" in Barbados. Such an explanation must include mechanisms for daytime suppression and nighttime uplifting of moist air. Garstang's research on the subject of flow over tropical islands led him to suggest the following mechanisms for such a situation.

During the daytime, the air near the surface of the island is heated by the warm earth beneath it and is, therefore, lighter than the cool air arriving with the northeast trade winds. Under the influence of these prevailing winds the lower, warmer, lighter air is able to move westward at a faster rate than the cooler, relatively heavy air above it. Daytime suppression occurs when the cooler air descends to fill the space left by the rapid westward movement of the lower air. At night the process occurs in reverse. The lower air is cooler and denser, hence slower moving. The upper air, on the other hand, is able to continue its westward movement at a less restricted rate. The relatively rapid removal of air aloft causes the lower air to be uplifted. It is this process that offers the necessary mechanism for the air to rise, condense, form cloud and ultimately cause precipitation to fall on the western side of the island at night.

Garstang's (1966) studies indicate several features of interest.

1. Maximum wind speeds occur over the ocean at night, with minimum speeds during the day.

2. The diurnal distribution of sensible heat exchange from the ocean to the atmosphere experiences maximum instability and turbulent exchange at night and a minimum during the early afternoon.

3. Higher values of easterly momentum are transferred downward during periods of maximum turbulent exchange (i.e., at night), with lower values being transferred downward during periods of minimum turbulent exchange (i.e., during the day). This pattern was observed at a station on the east coast of Barbados.

4. Study of the diurnal distribution of wind speed at central and west coast stations on Barbados shows maximum wind speeds occurring during the day and minimum wind speeds at night. This is the opposite of the east coast situation.

5. Given the difference between east and west coast stations, low level divergence must prevail over the island during the day and low level convergence must occur at night. (When tested using a mathematical model, maximum convergence was found to take place between 0200 and 0300 hours, with maximum divergence between 1500 and 1600 hours).

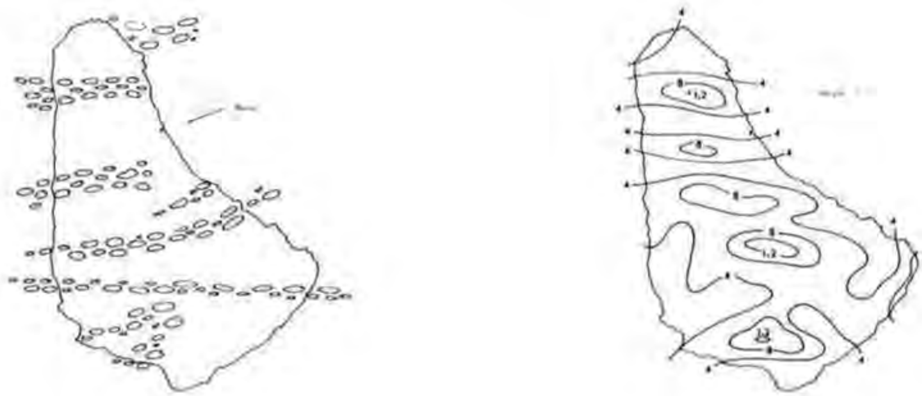
6. When mean summertime rainfall which occurs during periods of undisturbed synoptic conditions is plotted against divergence over Barbados, excellent agreement is obtained to support the relationship between divergence, subsidence, and the restriction of rainfall. It is the spatial pattern of this relationship that is of particular relevance to the study of rain bands. Garstang observed that: "...if significant subsidence occurs over the island during the day in response to low level divergence, this does not occur in a uniform fashion, but in a series of longitudinal rolls, lying parallel to the wind field".

It is possible that this longitudinal pattern of subsidence is partly responsible for the cloud "streets" which have been observed over the island both during the day and at night. Cloud "streets" are bands of cloud separated by clear sky. A typical cloud "street" pattern (corresponding in this case to the undisturbed synoptic conditions of April 7, 1963) is illustrated by figure 3. The probable distribution of rainfall for the same conditions are also shown in figure 3. It is important to note that although cloud "streets" may be observed throughout both the day and night, rain bands are predominantly a nocturnal phenomenon.

The Relationship Between Radiation Balances and Rainfall Patterns in Barbados

Garstang's work is interesting in that it examines possible mechanisms for both the creation of cloud "streets" over tropical islands during the day and the production of rain at night. Unfortunately, his work cannot entirely explain Barbados' nocturnal rain bands, for the following reason. If one accepts the suggestion that cloud "streets" may be caused by patterns of suppression (which Garstang believes occur over tropical islands only during the day), why, then do cloud "streets" and rain bands occur over western Barbados at night?

It is at this point that Garnier's work on the radiation balance of east-facing slopes may be relevant to the topic of rainfall patterns in Barbados. Garstang's research applies to tropical islands with flat topography. His work in Barbados was conducted in the area of the Vale, which is the island's flattest cross-section. It is suggested that while Garstang's description of the effects of heated islands must apply to the tropical island of Barbados, they cannot be assumed to act with equal intensity in all locations. For example, the wind that blows over the gentle, sugar cane-covered terraces of the Vale during a morning of undisturbed synoptic conditions must be less disturbed than the wind that traverses the hotter,



(a) cloud streets

(b) Percentage of stations recording rainfall

Fig. 3. Cloud Streets and Rainfall Distribution over Barbados under Undisturbed Synoptic Conditions (after Tyson, et.al., 1973).

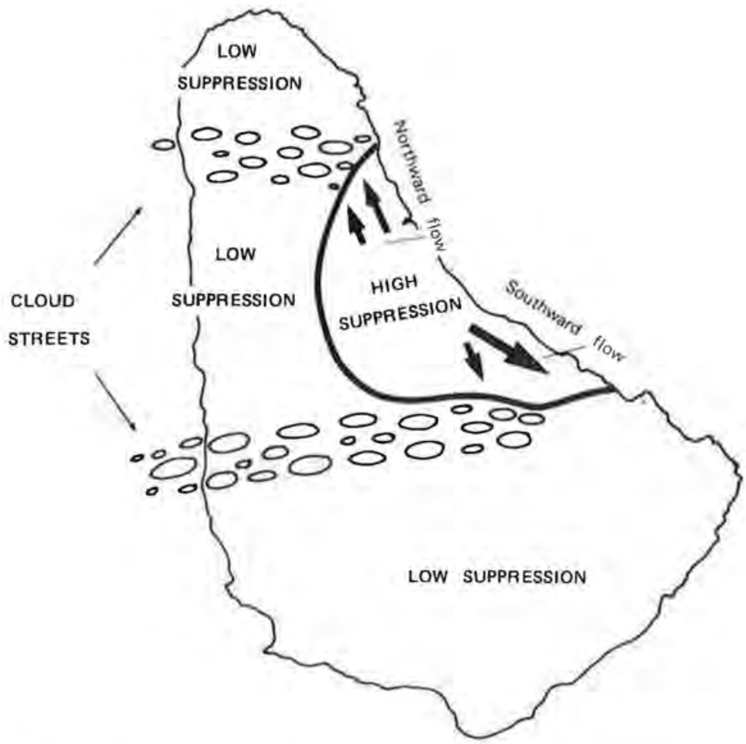


Fig. 4. Cloud Street Formation resulting from Differential Suppression.

rougher, unvegetated cliffs and gullies of the Scotland district and its surroundings. It is the temperature pattern and three dimensional wind field that evolve out of this differential heating and vertical displacement of the trade winds which must ultimately influence the pattern of cloud development during the day.

The most likely manifestation of this influence is the encouragement of north-south flows of air in response to varying degrees of suppression. The hotter the air just above the surface of the island, the greater will be its speed of western movement when impelled by the force of the trade winds. The greater this movement, the more intense must be the daytime suppression over that area. Applying this reasoning, the Scotland district and the area of predominantly east-facing slopes immediately surrounding it should establish a strong suppression regime during the period of intense heating (between 0700 and 0830 hours) observed by Garnier (1975). This strong suppression will contrast with that of the flatter, cooler areas to the north and south of the east coast study area. The areas to the north and south, being areas of less suppression, must provide outlets for the heated air of the Scotland district. As the Scotland district is amphitheatre-shaped, any north or south migration of warm air in response to areas of reduced suppression would result in orographic uplift, cooling and cloud formation. By the time the condensation level had been reached, the air would be under the influence of the less topographically-restricted levels of the trade winds, hence cloud "streets" would form parallel to the prevailing northeasterlies. One would expect, then, to observe one cloud "street" immediately to the north, and one immediately to the south of the east coast study area. This was, in fact the pattern observed by Garnier in response to the local "hot spot" described earlier.

The possible influence of the standard orographic model was intentionally downplayed earlier in this work in order to emphasize the point that it alone cannot explain Barbados' rain bands. Having elaborated on the less conventional forces which may influence cloud "street" and rain band formation, we must now attempt to integrate orographic considerations into the overall picture.

Garnier has described an area of the east coast that heats up rapidly in the early morning and remains warmer than other parts of the island until about mid-morning. Garstang's theory suggests that warm air from this region will move westward quickly, creating a suppressive downflow of the cooler prevailing northeast winds. What Garstang's model does not account for, however, is that, for this particular section of Barbados, westward movement entails orographic uplift over the central uplands. Once over the highlands, this warm air will encounter an area of weaker suppression. (Less intense suppression must occur in this area because of its lower morning temperature and the blocking by the central uplands of part of the trade wind momentum necessary to induce the lower warm air to move westward and be replaced by upper, cooler air). The "heat bubble" from the Scotland district must, upon encountering a region of reduced suppression, continue to rise. Cloud would be formed when the moist air cooled and condensed. The establishment of a linear cloud pattern at this point may well then be associated with the overall island divergence/suppression model described by Garstang.

Having examined the possible influence of the differential heating of east-facing slopes on the patterning of cloud "streets" during the day, one is still left with the task of explaining their survival throughout the night, as well as their nocturnal production of rain.

It is possible that, under undisturbed synoptic conditions, the mechanisms which are responsible for the patterning of clouds at the east coast would continue to function throughout the night. Two of their three major components (trade wind momentum and orographic uplife) remain constant or are increased at night. Garstang observed that east coast winds become stronger at night, thus a suppression regime that was already functioning would be intensified by the greater trade wind momentum. Orographic considerations remain constant. The third variable, temperature near the surface of the land, is thus the only factor which could reverse the process under calm synoptic conditions. Garnier observed that sunset does <u>not</u> induce selective cooling of the different physiographic regions. The north and south flows from the Scotland district would, therefore, not be expected to reverse at night.

It is empirically obvious that cloud "streets" survive throughout the night in Barbados and that they are responsible for the island's rain bands. An explanation for the fact that they drop more rain at night than during the day has yet to be found, however. One possibility is that radiative cooling of the cloud tops is sufficient to induce further instability within the clouds. The answers to this and other questions concerning Barbados' rain bands will only be found if a holistic approach to the problem is taken. Micro-, meso- and macro-climatological influences and their interrelationships must all be integrated into any model that hopes to explain the relationship between radiation balances and rainfall patterns in Barbados.

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SOME THOUGHTS ON THE CLIMATOLOGY OF SOLAR ENERGY POTENTIAL*

by

B.J. Garnier**

A consideration of the climatology of solar energy potential is an exercise in applied climatology in which one element of climate - solar radiation - is examined from the viewpoint of the factors governing the flux of energy on some form of solar collector. Such an examination involves looking at the solar radiation ($K\downarrow$) falling on surfaces of different gradient and azimuth in such a way as to discover the optimum exposures for solar energy collectors in different latitudes and climatic regions. It also involves assessing how much energy is, in fact, available during different seasons and how it is distributed over short time periods within those seasons.

In the present context the solar energy collector in question will be taken to be a flat-plate collector, placed in a position of fixed azimuth and gradient. The general radiative income of such a collector will be controlled by those factors which determine the general solar radiation of any part of the earth at a given time: earth/sun relationships and the physical state of the atmosphere expressed mainly through the weather. The final control of the income, however, will be governed by the collector's azimuth and gradient. In other words, the actual income of the collector will, in the last analysis, be strongly influenced by the way the surface is

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** B.J. Garnier is Professor of Climatology in the Department of Geography at McGill University.

exposed. As a result, its income may be significantly different from that obtained from observations recorded in the usual way on a horizontal surface.

By comparison to that of a horizontal surface, the response of a sloping surface to incoming solar energy, is more sensitive to direct solar radiation than it is to sky-diffuse radiation. This is because the former comes from a point source in the sky, whereas the latter comes from numerous parts of the sky. Indeed, in evaluating daily totals of solar radiation on a slope little error is introduced by assuming that sky-diffuse radiation arrives equally from all parts of the sky, i.e. that it is isotropic. There are two reasons for this: firstly, differences arising from different sun elevations tend to be cancelled out when integrated over daylight hours (Robinson, 1966, p.122); and, secondly, detailed variations in the arrival of sky-diffuse radiation are greatest under clear sky conditions when the contribution of sky-diffuse radiation to total solar radiation income is least, and they are minimal under overcast or very cloudy conditions when the contribution of sky-diffuse radiation to the total solar radiation income is greatest.

If sky-diffuse radiation is taken to be isotropic, the total falling on a slope by comparison with that on the horizontal is given by the expression (Kondratyev, 1965, p.332):

$$D_s = D_h \cos^2(\theta/2) \qquad (1)$$

where D_h is sky-diffuse radiation on the horizontal and θ is the gradient of the slope. From this it follows that the daily total of sky-diffuse solar radiation on a slope will always be less than that on the horizontal up to a maximum difference of 50% when θ=90°.

By contrast, the flux of direct solar radiation on a slope will frequently be substantially greater than on the horizontal, especially under low sun conditions. A convenient way to make the relevant calculations can be derived from the expression (Garnier and Ohmura, 1968, 1970):

$$I_s = I_m \cos(\vec{X}_{\wedge}\vec{S}) \qquad (2)$$

in which I_m is the flux per minute of direct solar radiation falling on a surface normal to the sun's rays, $\vec{X}$ and $\vec{S}$ are unit co-ordinate vectors describing the gradient of the slope and the position of the sun in the sky respectively, and the symbol $\wedge$ denotes the angle between the two vectors.

It has been shown elsewhere (Garnier and Ohmura, 1968) that, for a given moment of time,

$$\cos(\vec{X}_{\wedge}\vec{S}) = [(\sin\phi \cos H)(-\cos A \sin Z_x) - \sin H (\sin A \cos Z_x)$$

$$+ (\cos \phi \cos H) \cos Z_x] \cos \delta$$
$$+ [\cos \phi (\cos A \sin Z_x) + (\sin \phi \cos Z_x)] \sin \delta$$

where ϕ is latitude, δ is solar declination, H is the hour angle measured in degrees from noon (negative to the east), A is the slope azimuth, and Z_x is the slope gradient.

Equation (2) may be used to obtain totals of direct solar radiation on a slope by summation over appropriate time intervals. Ohmura (1969) has shown that a time interval of 30 minutes will give an accuracy to within 5 per cent of that using calculations at one-minute intervals for slopes up to 90°. Since this degree of accuracy is within the observational error of commonly-used instruments for observing solar radiation (Latimer, 1971), such an interval can normally be used for solar energy purposes. Indeed, a time interval of one hour is probably sufficient. Tests have shown that the system provides an accuracy of ±5% for daily totals of solar radiation on a slope when used in conjunction with the evaluation of sky-diffuse radiation given in equation (1).

The calculation of direct solar radiation on a slope (S_s) may, in practice, proceed in one of two ways: by the use of observations of direct solar radiation made at short time intervals (one hour or less), or by the use of an average transmissivity for the day (p) derived from daily totals of observed direct radiation. In the former case the appropriate value of I_m in equation (2) may be obtained from the expression

$$I_m = I_h/\cos Z$$

where I_h is the observed value on the horizontal and Z is the zenith angle of the sun obtained from:

$$\cos Z = \sin \phi \sin \delta + \cos \phi \cos \delta \cos H.$$

The alternative method, employing daily totals, is to use the daily total to establish a "mean transmissivity" for the day. This "simple value of transmission" (Haltiner and Martin, 1957) can then be incorporated in equation (2) by means of:

$$I_m = I_r \, p^m$$

where I_r is the flux per minute of extraterrestrial radiation for the date in question, and m is the optical air mass given by:

$$m = \sec Z = 1/(\sin \phi \sin \delta + \cos \phi \cos \delta \cos H)$$

under the condition that $Z \leq 70°$. Otherwise the value of m must be obtained from tables (List, 1963). By these means a daily total of direct solar radiation on a slope (S_s) may be obtained by summation from sunrise on the slope (t_1) to sunset (t_2) in the form:

$$S_s = I_r \sum_{t=t_1}^{t_2} p^m \cos (\vec{X}_\wedge \vec{S}) \Delta t \qquad (3)$$

where Δt = 30 minutes.

In theory it would seem that the use of observations at short time intervals of an hour or less would, if available, be preferable to using an average transmissivity for the day. The flux of direct solar radiation on a slope is sensitive to changes in atmospheric transmissivity. An average value assumes uniformity throughout the day. This is by no means always the case. A change in weather during the morning or afternoon, therefore, can be expected to have noticeable consequences for short-wave radiation intensities and these consequences should be differentially reflected in the radiation totals of slopes of contrasting slope or azimuth. So far as solar energy is concerned, however, these potential differences through the day do not seem to have a major effect on resulting daily totals. Table One, for example, compares the daily totals derived from using the two approaches described above for different periods at Toronto and Goose Bay. The data suggest that the discrepancy arising from the two procedures is minimal for an azimuth of solar noon (180°) and only becomes noticeable at azimuths in the vicinity of sunrise or sunset. Evidently the consequences of changes in atmospheric transmissivity through the day which are, of course, reflected in the daily total of direct radiation from which the mean transmissivity is derived, are averaged out for slopes with an azimuth near solar noon. Such an azimuth is usual for a flat-plate collector. For solar energy, therefore, it would appear that daily totals of direct solar radiation observed on the horizontal can be safely used to evaluate the flux of energy from the direct beam on a slope.

Tables Two and Three show the results of calculations for different surfaces on different dates at two locations in Canada. The data in the first of the two tables were prepared in connection with the potential for solar heating of a mountain chalet (Garnier, 1977). The second table refers to conditions in the Toronto urban area. Both tables tell essentially the same story. They show that relying only on horizontal observations as an indicator of solar energy potential is unreliable. At both places there was high solar energy potential on the day of lowest solar radiation on a

TABLE ONE

Correlation Coefficients between Direct Solar Radiation on a Slope Calculated from Daily Totals (X) and Hourly Observations (Y)

Place	Slope Gradient	Group	No. of Obs.	Azimuth 60°	120°	180°	240°	300°
Toronto	60°	1	9	.83	.90	1.00	.93	.87
		2	15	.76	.85	1.00	.91	.87
		3	17	.86	.90	1.00	.90	.80
		All	41	.90	.87	.99	.91	.92
Goose Bay	60°	1	6	.99	.99	1.00	.99	.97
		2	9	.48	.82	1.00	.95	.85
		3	10	.70	.80	.99	.89	.87
		All	25	.83	.87	.99	.93	.92

Group 1: Low sun period
Group 2: Medium sun period
Group 3: High sun period

horizontal surface. The low-sun days of winter are particularly noteworthy in that the solar radiation on a flat-plate collector would have been more than double that measured on the horizontal. In each table, information has been added to show the power produced by the flux of solar radiation on a flat-plate collector of 20 m^2 area. In both cases the maximum power during daylight hours was in winter. The challenge is, of course, to be able to use this power with an efficiency of both conversion and storage sufficient for it to provide useful energy in a practical way. In the present context, it is important to note that much higher values of potential solar energy can be shown to exist than is revealed by relying solely on horizontal totals of global (K↓) values. The reason for this discrepancy lies in the important role played by direct radiation in the total energy received by sloping surfaces.

A further feature of the situation is that the discrepancy between slope and horizontal surface values tends to be greatest under low sun conditions, which is the very season when heating requirements are normally greatest. The seasonal pattern of contrast is shown in figure 1. The calculations for this diagram were based on dividing the year into periods of "equal" solar declination. These are periods of unequal length

TABLE TWO

Solar Heating of South Facing Roofs at Deer Creek, Alberta: 53°N

	February 5, 1970			April 9, 1970		
	Horiz.	Roof	Slope	Horiz.	Roof	Slope
		70°	50°		70°	50°
Direct (S) (Langleys)	160	460	427	264	306	339
Global (K↓) (Langleys)	199	485	459	446	429	489
$K{\downarrow}_s/K{\downarrow}_h$	–	2.44	2.31	–	0.96	1.09
KW*	5.76	12.56	11.88	7.80	7.50	8.54

* Power on Flat-plate collector area $20m^2$ averaged over daylight hours.

TABLE THREE

Solar Heating of a South Facing Slope (50° gradient) at Toronto, lat. 44°N

Date	K↓ (langleys)		BTU($ft^{-2}day^{-1}$)		KW*
1972	Horiz.	Slope	Horiz.	Slope	
Jan. 14	205	421	755	1551	10.82
Apr. 12	457	426	1684	1570	7.56
May 24	300	247	1106	910	3.88
Oct. 11	330	462	1216	1703	9.66

* Power on Flat-plate collector area $20m^2$ averaged over daylight hours.

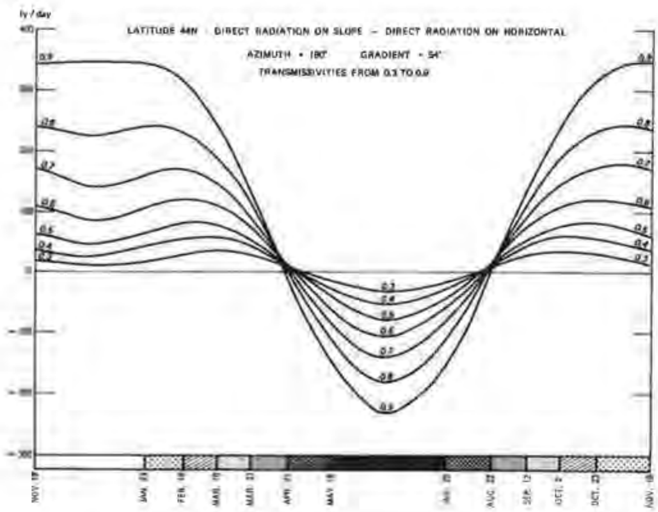
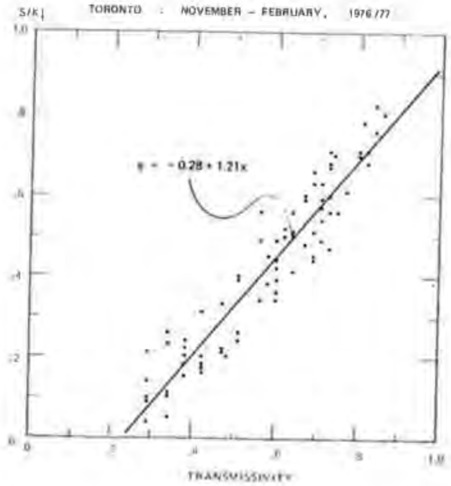


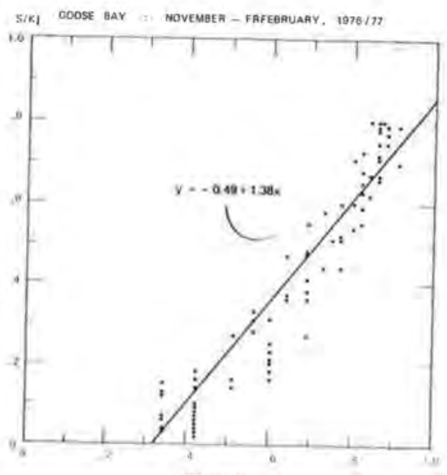
Fig. 1. Seasonal Contrasts in Direct Radiation on a Slope and on the Horizontal under Different Transmissivities.

during which the use of a "central value" of solar declination will produce values within the 5% degree of accuracy referred to earlier. The concept has been explained in more detail elsewhere (Garnier and Ohmura, 1970). The diagram makes it clear that on the "low sun" side of the equinoxes a flat-plate collector at Toronto, suitably positioned, can be expected to receive noticeably more direct solar radiation than a horizontal surface down to quite low atmospheric transmissivities. The excess of slope over horizontal is particularly noticeable when the atmospheric transmissivity is 0.6 or greater. During the winter period, the excess at a value of $p = 0.6$ is approximately 100 langleys per day. This converts to about 2.6KW of power averaged over daylight hours on a flat-plate collector exposed at an angle of 54° (the latitude of Toronto plus 10°).

It must be clear by now that a key to the value of solar energy potential is to be found in the value of the atmospheric transmissivity of a given day. It follows that a climatology of solar energy potential should devote attention to this key value. Equation (3) has already made the importance of atmospheric transmissivity clear in respect of calculating daily totals of direct radiation on a slope. It should be noted, however, that atmospheric transmissivity can also be used as an empirical means for



(a) Toronto



(b) Goose Bay

Fig. 2. Linear Regressions between Transmissivity at the ratio Direct (S) and Global (K↓) Radiation.

evaluating K↓ on a horizontal surface. This makes it possible to estimate the total solar energy on a slope with the aid of equation (1).

Linear regression analyses relating atmospheric transmissivity (X) to the ratio of direct (S) to global (K↓) radiation (Y) for the winter period of 1976/77 for both Toronto and Goose Bay are illustrated in figure 2. The correlation values for the relationships at the two locations were: for Goose Bay, $r^2 = 0.92$ and $s_{y.x} = 0.08$, and for Toronto $r^2 = 0.91$ and $s_{y.x} = 0.07$. These are high correlations. The same degree of correspondance was found at other seasons of the year for the two places, with only small variations in the values of the regression constants which are shown in figure 2. From relationships such as these it is a simple matter to estimate the value of sky-diffuse radiation. For example, the low-sun value of K↓ at Toronto obtained from the regression equation in figure 2 will be:

$$K\downarrow = S/(-0.28 + 1.21X)$$

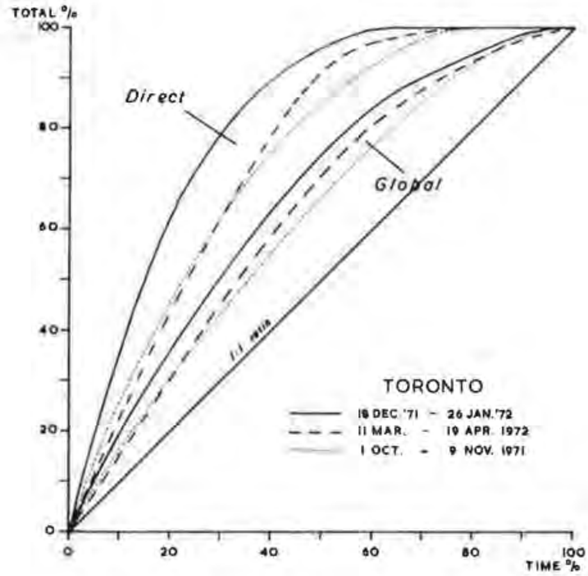
where S is the value of direct radiation calculated for a given transmissivity (X). From this it follows that

$$D = K\downarrow - S.$$

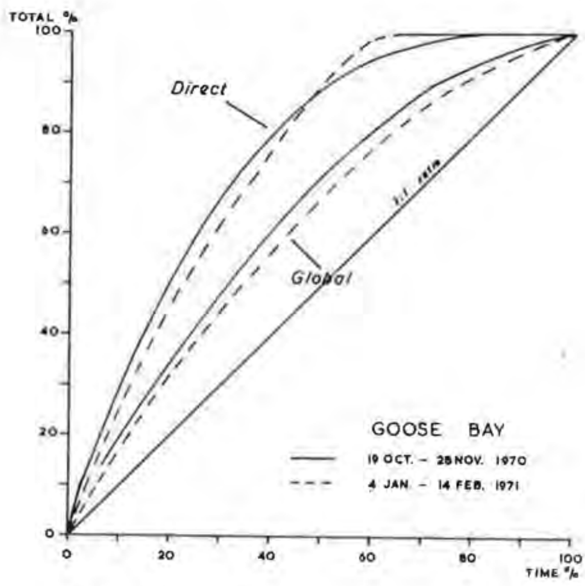
One matter to which no attention has yet been paid is the question of the frequency of oocurrence of "useful days" from the solar energy point of view. This is clearly a matter of considerable concern to solar energy engineers who must develop systems which both adequately use and adequately store solar energy against sequences of sunless or "low value" days. Once again the use of K↓ on a horizontal surface should be adopted with caution for this purpose. This is illustrated in figure 3.

It can be seen from this figure that at both Toronto and Goose Bay for the time periods studied the total accumulation of direct solar radiation was achieved over a relatively shorter time period than the total accumulation of K↓. Indeed, at both places during winter some 80% of the former was achieved in less than 50% of the time, and all direct solar radiation accumulated in approximately 60% of the time. The accumulation of K↓, however, was much more evenly spread.

In view of the importance of direct solar radiation to the totals received by a flat-plate collector, these differences are significant for any reliable use of solar energy as a form of power. The contrast between the situation on a slope and on the horizontal are further illustrated in figure 4. Both in a relative sense and in terms of actual values, the accumulation of solar energy totals on a slope are concentrated into a



(a) Toronto



(b) Goose Bay

Fig. 3. Ranked Cumulative Totals of Direct and Global Radiation on a Horizontal Surface at Toronto and Goose Bay.

TABLE FOUR

Table of Estimated Daily Totals of Solar Energy falling on a flat plate collector at Toronto for different transmissivities and "equal" solar declination periods.

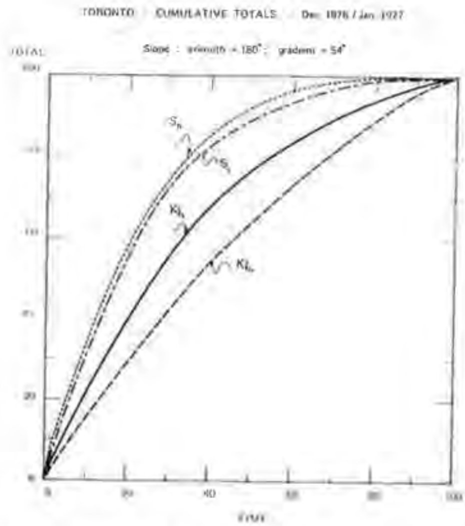
Collector azimuth = 180°, gradient = 54°

	Period	Transmissivities				
		0.5	0.6	0.7	0.8	0.9
langleys per day	1	123	182	267	386	559
	2	258	326	417	533	706
	3	309	388	480	588	718
	4	408	479	557	645	744
	5	493	549	609	672	739
	6	463	517	543	617	663
	7	490	527	559	586	608
Kilowatts*	1	3.3	4.9	7.1	10.3	14.9
	2	6.1	7.7	9.8	12.5	16.6
	3	6.5	8.2	10.2	12.5	15.2
	4	7.9	9.3	10.8	12.5	14.4
	5	8.8	9.8	10.9	12.0	13.2
	6	7.6	8.5	9.0	10.2	10.9
	7	7.5	8.0	8.5	8.9	9.2

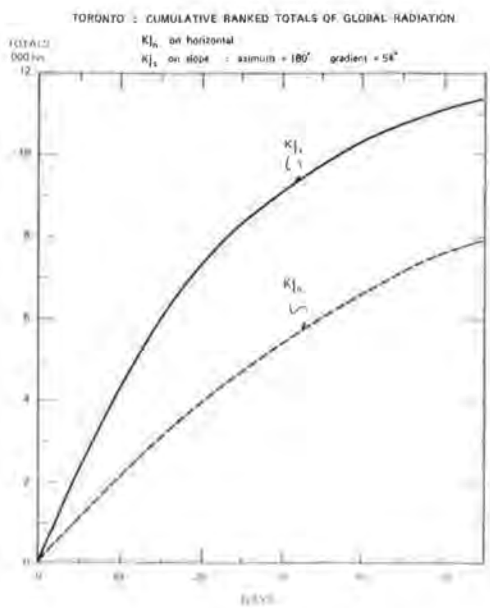
*Solar power falling on a flat plate collector area $20m^2$, averaged over daylight hours.

shorter period than on the horizontal. In other words, the relatively high values of daily solar energy referred to earlier are features of a relatively limited number of days. Both their evaluation and prediction are closely related to a good understanding of the occurrence and forecasting of daily atmospheric transmissivity.

The thrust of the argument contained in these thoughts on the climatology of solar radiation potential has been towards stressing the importance of a good evaluation and prediction of an average daily value of atmospheric transmissivity. The usefulness of such a value is illustrated in Table Four. To the extent that the precedures discussed earlier are valid, the table shows the power available on a flat-plate collector with an area of 20 m^2 under different atmospheric transmissivities at different times of year at Toronto. The amount of solar power received by such a collector is considerable. It is certainly well in excess of home heating requirements. Bolton, in an unpublished manuscript (Bolton, n.d.) indicates that in Ottawa an average home uses about 0.5 KW averaged over 24 hours per day averaged over the year. Hoffman (1975) indicates that in January, at Vancouver, an



(a) Percentage of Total against Percentage of Total Time.
S = Direct Radiation; K↓ = Global Radiation
Subscripts refer to horizontal (h) and slope (s).



(b) Total Accumulation of Global Radiation Ranked Against Number of days of Record for a Slope and on the Horizontal.

Fig. 4. Ranked Cumulative Totals of Solar Radiation on a Slope at Toronto.

average of 0.71 KW averaged over a day is needed. Allen (1975) suggests a daily requirement of 0.87 KW at Gananoque in January. The figures provided by Table Four are commensurate with these requirements at about a 10 - 15% efficiency utilising the solar power on days on which atmospheric transmissivity is ≥ 0.5. In the winter half of the year for 1976/77 at Toronto this condition occurred on about 45% of total days.

The challenge to solar energy engineers is how to make use of and store the solar power available. For the climatologist, the challenge is to provide a meaningful analysis which provides the solar energy engineer with reasonably reliable information concerning the solar energy they have to work with. It is suggested that this may be done by developing a synoptic climatology, together with an analysis and prediction of weather conditions, centred upon an improved understanding of the manifestation and frequency of occurrence of different levels of daily atmospheric transmissivity in different parts of the world.

Acknowledgement

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NEWS AND COMMENTS

The non-organization known as Friends of Climatology slumbered somewhat in 1979. With the tenth anniversary of its non-birth occurring this year, it was appropriate for the Friends to meet at the University of Windsor, April 18 and 19, 1980 under the auspices of Marie Sanderson, where the first gathering was held in 1970. There was a good attendance of some fifty persons, including a dozen students, at an evening session on the Friday and a morning session on the Saturday. The general theme of the meeting was the World Climate Program and the Canadian contribution to it through a group of committees, with the AES climatic unit as a core. There was also discussion of the role of the Friends in the program. Lead speakers were Ken Hare, Morley Thomas, and Gord McKay, but most of those present participated in one way or another from the floor.

To ensure against further slumberings several decisions were taken. An "activity committee" of four was set up, Ken Hare was appointed Orbiting Climatologist for 1980/81, Ben Garnier was asked to try to infiltrate the Learned Society Conference meeting, to promote climatological work and invitations for the annual non-meetings for the next two years were obtained: at Chalk River for 1981, and at McMaster University for 1982.

The University of Delaware has established a PhD Program in Applied Science-Climatology. This will be an interdisciplinary programme administered by the Department of Geography (Chairman John R. Mather). Entry to the program will be normally by way of a master's degree, with climatology or meteorology as the central concentration. For the Doctoral program, students will be required to obtain in-depth knowledge in two substantive areas within a a broad overview of climatology: one of the in-depth areas must be topical, e.g. bioclimatology, urban climatology or microclimatology, and the other must be methodological such as statistical methods, mathematics, or computer science. Each student's program, however, will be individually tailored to the student's interests.

Climatologists will be interested in the arrival of two new journals in the publication market. One is the Journal of Climatology to be published quarterly, starting in January, 1981, by John Wiley and Sons Inc. The editor is Professor S. Gregory of the Department of Geography, University of Sheffield, England. Specimen copies and further information on the journal may be obtained from the publishers.

A second journal of interest has already appeared. It is Physical Geography, edited jointly by Professor Antony R. Orme of UCLA Department of Geography, and Professor John E. Oliver, Department of Geography and Geology at Indiana State University, Terre Haute. Publication of the journal is by W.H. Winston and Sons, 7961 Eastern Avenue, Silver Spring, Md., 20910, to whom inquiries and subscriptions, varying from \$16.50 to \$29.00 according to individual status, should be sent.

The recent annual meeting of the Association of American Geographers, held at Louisville, Kentucky, from April 13-16, 1980, witnessed a strong resurgence of climatology. Stirrings in this direction had occurred in the 75th anniversary meeting held in Philadelphia in April 1979. These stirrings bore fruit in Louisville with the official recognition by the Association of climatology as a special interest group. It also resulted in a total of some ten sessions devoted specifically to various aspects of climatology. The increasing importance of our discipline was also noticeable in the number of climate-oriented papers in other sessions, notably remote sensing, geomorphology and biogeography. A Canadian who was particularly active in all this climatological resurgence was Marie Sanderson of the University of Windsor, both as a sessional chairman and discussant.

The 14th annual congress of the Canadian Meteorological and Oceanographic Society (CMOS) will be held in Toronto, May 22-27, in conjunction with the annual meeting of the American Geophysical Union. Next year's CMOS annual congress (1981) will be held at Saskatoon from May 26-29.

In March of last year the annual meeting and a workshop of the Alberta Climatological Association were held in Edmonton. The proceedings, compiled by J.M. Powell, have now been published as information report NOR-X-217 by the Northern Forest Research Centre of Environment Canada and the Resource Evaluation and Planning Division of Alberta Energy and Natural Resources. The theme of the publication is Socioeconomic Impacts of Climate, and sections of the report cover Agriculture and Forestry, Industrial Applications-Energy, Recreation, and Urban and Regional Planning.

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