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Publication de la Société canadienne de météorologie et d'océanographie, le Bulletin climatologique offre un moyen d'information sur la climatologie. Le comité de rédaction encourage en particulier la soumission de manuscrits sur la climatologie appliquée (comme l'agriculture, le commerce, l'énergie, l'environnement, la pêcherie, la sylviculture, la santé, les loisirs, les transports, et les ressources en eau), les changements et la variabilité du climat, la prospective climatologique, les applications des modèles du climat (inclus la climatologie physique), et les études régional (inclus les océans). Il est publié grâce à une subvention accordée par le gouvernement canadien par l'intermédiare du Conseil de recherches en sciences naturelles et en génie.

Les auteurs peuvent choisir de soumettre leurs manuscrits aux "Articles", "Notes de Recherches", ou "Nouvelles et Commentaires". Ils doivent l'indiquer sur la lettre d'accompagnement du manuscrit. Les articles de recherche et les "Notes" sont indépendamment soumis à l'examen d'au moins deux appréciateurs anonymes. Le rédacteur en chef examine les "Nouvelles et Commentaires" conjointement avec le comité de rédaction. On accepte les articles soit en français, soit en anglais. Il faut envoyer un résumé, de préférence en français et en anglais.

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Foreword / Avant-Propos

At its annual general meeting at the Victoria Congress in May, CMOS decided to investigate the possibility of a new vehicle for its publishing activity in the domain of climate. *Climatological Bulletin* will continue in its present form until the end of 1991, having served as this vehicle since CMOS took it over in 1983. Increased interest in the *Bulletin* over the past two years on the part of both authors and subscribers may well indicate that the time is ripe for the introduction of a larger, upgraded Canadian journal of climate.

A l'Assemblée générale annuelle au Congrès de Victoria en mai, la SCMO a décidé de rechercher la question d'un changement pour ce qui est de son activité de publication dans le domaine du climat. Le *Bulletin climatologique* existera encore sous sa forme actuelle jusqu'à la fin de l'année 1991. Il répond aux besoins de la Société depuis 1983. Cependant, l'intérêt d'auteurs et d'abonnés dans le *Bulletin*, qui a beaucoup augmenté les deux dernières années, signale peut-être l'arrivée du moment de lancer une nouvelle revue canadienne du climat, agrandie et améliorée.

Alec Paul
Editor/Rédacteur en chef

Some Observations on the Effect of Chinooks on Field Microclimates and Soil Moisture Status in Southern Alberta

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[Original manuscript received 14 September 1988;
in revised form 20 December 1989]

ABSTRACT

Microclimatic observations were made on a fallow plot and a plot with winter wheat stubble (about 48 cm high) during two periods of late fall chinook weather in southern Alberta. Only slight changes in near-surface microclimates were noted in response to chinook conditions when the soil surface was dry. Changes in net radiation were attributed to the differences in albedo of the two surfaces. There were only slight changes in evaporation and hence soil water. The difference in evaporation rates was attributed to availability of soil water and the effects of the surface cover. However, the near-surface microclimate of the two surfaces was significantly different in response to chinook conditions following snowfall accumulation on the respective surfaces and the subsequent disposition of energy and moisture. Rapid ablation of snowcover on a fallow surface allowed little of the moisture to be added to the soil. The capacity of the stubble surface to accumulate significantly more snowfall than the fallow surface resulted in more energy being required for the subsequent evaporation, sublimation, melting and infiltration. Infiltration of meltwater provided a significant contribution to the soil moisture reserve of the unfrozen soil.

RÉSUMÉ

Nous avons observé les différences micro-climatiques de deux plans de terre, l'un en jachère et l'autre en chaume (blé d'hiver d'une hauteur de 48 cm), soumis à des épisodes du Chinook vers la fin de la saison automnale dans le sud de l'Alberta. Lorsque la surface du sol était sèche, nous n'avons noté que très peu de changements micro-climatiques près de la surface. La différence de radiation net a été attribué à la différence d'albédo entre les deux surfaces. Quant au taux d'évaporation et à la quantité d'eau disponible dans le sol, les changements étaient également minimes. La différence des taux d'évaporation a été attribué à la quantité d'eau disponible dans le sol et au type de couvert de la surface. Toutefois, les deux surfaces ont démontré des différences significatives sur le plan du transfert d'énergie et d'humidité durant un épisode du Chinook suite à une accumulation de neige. La fonte rapide de la neige sur le plan en jachère n'a permis qu'une addition minime d'eau dans le sol. Par contre, la capacité de la chaume à accumuler une plus grande quantité de neige a résulté en un besoin plus élevé d'énergie requise pour toute évaporation, sublimation, fonte et infiltration subséquente. L'infiltration des eaux de fonte a ajouté une contribution significative à la réserve d'eau sols non gelés.

1. INTRODUCTION

The climate of southern Alberta and Montana was often referred to in historical records (Mackenzie, 1801; McCaul, 1888; Thompson, 1962). The intensity and frequency of chinook conditions in Canada are greatest in the agricultural areas of southern Alberta (Longley, 1967; Grace 1987a). Chinook is the name applied to strong, warm, dry winds that sweep down the eastern slopes of the Rocky Mountains. A chinook, a type of foehn wind, is the result of the passage of a typical mid-latitude low-pressure center at the lee of the Rocky Mountains into which air is drawn. Often this is associated with a ridge of high pressure over the mountains (see Grace and Hobbs, 1986).

Little attention has been given to the influence of chinooks on agriculture. Other than during the short growing season (May-August), land that is used for crop production in this area is free of growing vegetation. The soil surface is either covered with stubble and crop residue from the most recent harvest, in keeping with minimum tillage practices, or the soil is free of surface cover as in the fallow stage of a conventional crop-fallow rotation system (Lindwall and Anderson, 1981; Lindwall, 1986).

Since crop growth normally removes all available moisture from the soil during each growing season, recropping dryland farming operations on the prairies depend on precipitation during the fall, winter and early spring to ensure adequate soil moisture reserves at planting time (Grace, 1987b). In fallow-rotation cropping systems, overwinter soil moisture accumulation is also important. It is during this important recharge time that soil moisture conservation practices are most critical. Several studies have attempted to measure or calculate the effect of chinook conditions on snow ablation (the removal of snow by sublimation, melting, evaporation and wind) (Ashwell and Marsh, 1967; Golding, 1978); however, little attention has been paid to the effect of chinook conditions on field microclimates and especially on soil moisture. Recent investigations (Lal and Steppuhn, 1980; Nicholaichuk *et al.*, 1985; Nicholaichuk, 1986; Caprio *et al.*, 1986) have demonstrated various techniques of snow management to accumulate snowpack on fields during the winter and have indicated the applicability of these management strategies for the enhancement of soil moisture reserves for areas of the Canadian prairies that are not dominated by chinook conditions.

Although the average annual snowfall at Lethbridge is 141 cm, it is more common not to have snow on the ground during the winter months than it is to have snowcover, owing to the frequency and intensity of chinook conditions (Grace and Hobbs, 1986). Thus, that portion of the snowfall which can be manipulated for agricultural purposes is dependent upon the quality and distribution of the snowpack, the physical characteristics of the landscape, and the subsequent loss due to chinook conditions. The objective of this study was to examine the near-surface microclimates of the two most common field situations, stubble and fallow, under chinook conditions during the important soil moisture recharge period in the fall.

2. METHODS, MATERIALS AND SITE DESCRIPTION

The study was conducted at the Agriculture Canada Research Station, Lethbridge during the fall of 1987. The physical properties of the Orthic Dark Brown Chernozemic clay loam soil are presented in detail in Table 1.

TABLE 1. Soil physical properties of Lethbridge clay loam (Chernozemic Brown Orthic)

Depth interval (cm)	Mechanical composition (%)			Bulk density (g/cm^3)	Saturated hydraulic conductivity (cm/min)
	Sand	Silt	Clay		
0–20	43.92	27.12	28.96	1.26	0.012
20–45	40.92	29.92	29.16	1.28	0.026
45–75	38.92	29.92	31.16	1.32	0.020

The recommended strip farming cropping practice for southern Alberta (Anderson, 1966) is a field oriented at right angles to the prevailing westerly winds. The study was conducted in an open field on 40-m wide strips. The winter wheat was planted in September of 1986 and harvested in August 1987. Two 10 by 40-m sections of the strip received different surface treatments. In a manner consistent with summerfallow practices, one plot was not seeded at the time of planting and subsequently received three blade tillage treatments during the following growing season to control weed growth. This plot had very little crop residue on the surface during the observation period. The other plot had been planted with winter wheat and harvested in such a way as to leave tall standing stubble and some crop residue on the surface. The standing stubble (291 g dry matter per m^2) ranged from 45 to 50 cm in height with a mean of 48 cm. The amount of detached residue lying on the surface was 133 ± 5 g dry matter per m^2. Other areas of the strip were either left after harvest with short (10 to 20 cm high) standing stubble or received one blade tillage treatment.

The plots were instrumented for microclimatic measurements with Middleton CN1 pyrradiometers for measurement of net radiation, Kipp and Zonen pyranometers for measurements of incoming and reflected solar radiation, and three Thornthwaite soil heat flux sensors in each plot at a depth of 5 cm for measurement of soil heat flow. A Campbell Scientific model 201 temperature and relative humidity probe was used for measurements of atmospheric humidity and temperature. Met-One model 013 anemometers were employed in the measurement of wind speed at a height of 1.5 m. Air temperature profiles in each of the plots were measured with shielded copper-constantan thermocouples at heights of 5, 10, 50, and 150 cm. Soil temperatures were monitored at depths of 2, 5, 10, 25, and 50 cm in each of the plots with networks of 10 copper-constantan thermocouples wired in parallel and placed at each depth interval. Soil surface temperatures were measured with a network of 10 copper-constantan thermocouples carefully positioned on the soil surface and shielded by lightly covering the thermocouples with a few millimeters of soil or crop residue depending on the surface configuration of the microsite. Signals from the sensors were monitored and recorded with a Campbell Scientific CR7 datalogger.

Evaporation was measured using *in situ* weighing lysimeters containing undisturbed soil cores. The weighing lysimeters were 50 cm in diameter and 50 cm deep. Standardized, sealed, temperature-compensated, beam type load cells (Interface model SSB-500) were used for monitoring the weight of the lysimeters. The lysimeters were calibrated against standard weights to ± 0.1 mm water equivalent. Soil moisture was monitored using neutron attenuation from a series of five access tubes in each plot, standardized against gravimetric measurements. Special attention was given to gravimetric sampling in the near-surface layer, the top 30 cm, where the accuracy of neutron attenuation techniques is less than at greater depths.

3. RESULTS AND DISCUSSION

Two time periods of fall chinook conditions are discussed in detail here. The first occurred on November 10–13, under conditions of a dry soil surface. The second time period examined, November 16–22, followed a snowfall.

a) *November 10 to 13 1987*

The presence of a chinook arch, relatively warm temperatures, large vapor pressure deficits, and a strong west wind from November 10 to 13 1987 was typical of fall chinook conditions in southern Alberta (Grace and Hobbs, 1986). Air temperatures at a height of 1.5 m ranged from 8 to 14°C during most of the period. Windspeeds recorded at a height of 1.5 m ranged from 4 m/s to a maximum of 10.9 m/s. Average hourly windspeeds of 10 to 18 m/s with gusts as high as 22 m/s were recorded at a height of 10 m at the nearby (1 km) Agriculture Canada Weather Observing Site. Relative humidity varied between 30 and 50% with vapor pressure deficits often approaching 1 kPa. The falling air temperatures, reduced vapor pressure deficits and declining windspeeds during the late evening hours of November 12 indicated the end of chinook conditions.

During the November 10 to 13 observation period the soil surface layers were relatively dry (Table 2) because of the lack of any recent precipitation. There was no snow cover on the ground at this time. The low values of soil moisture at all depths (0 to 130 cm) in the stubble plot were the result of crop water

TABLE 2. Percent soil moisture (v/v) from neutron attenuation for the stubble and fallow study plots, November 10, 1987 (means of measurements taken at 5 locations within each plot).

	Fallow		Stubble	
Depth (cm)	Mean	S.E.	Mean	S.E.
0–30	16.00	0.93	9.93	0.73
30–60	19.63	0.33	9.90	1.00
60–130	16.00	0.63	11.63	0.70

use during the past growing season. Owing to the dry surface, little evaporation was recorded during the 3-day period with the lysimeter on the stubble plot losing 0.2 mm of water and the lysimeter on the fallow plot losing 0.8 mm of water (Figure 1).

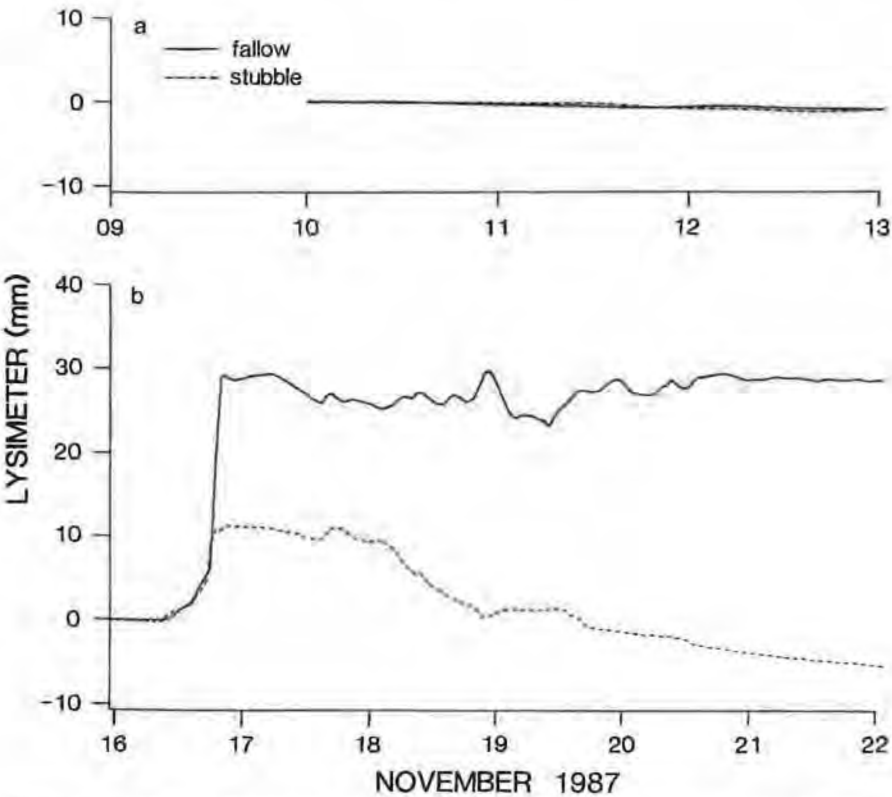

FIGURE 1. Mean hourly weight of weighing lysimeters (units of weight mm of water equivalent) for stubble and fallow plots for (a) November 10–13 and (b) November 16–22, 1987 with a 170 kg zero offset for the stubble lysimeter and a 225 kg zero offset for the fallow lysimeter.

Several factors combined to influence the evaporation rates from the two surfaces. There was more absorbed energy available for evaporation on the fallow surface than on the stubble plot owing to the increased albedo of the stubble (Figure 2a).

Complete energy balances for a clear day, November 12, for the fallow and stubble surfaces are displayed in Figures 2b and 2c. Values calculated for the albedo of both surfaces exhibited a diurnal variation with highest values observed during the early morning and evening hours and minimum values observed at midday. This was attributed to the change in the angle of solar incidence (Grace, 1975; Oke, 1978). Minimum midday (1000 to 1400 h) values of albedo for the bare fallow surface ranged between 15 and 20% whereas the values for the stubble surface varied between 25 and 38% and were within the range of values reported elsewhere for dry soil and vegetated surfaces (Oke, 1978). The greater albedo of the stubble plot resulted in a smaller amount of net radiation or absorbed energy than on the fallow plot (Figure 2b, 2c). During the midday hours of November 12 there was 22% greater net radiation on the fallow plot than on the stubble. The radiant energy balance of these surfaces was dominated by differences in shortwave energy absorption.

Another factor affecting the evaporation rates of the two plots was the difference in surface cover. The effect of the surface residue on the stubble plot was to increase the diffusive resistance for evaporation. This effect, combined with the reduced near-surface windspeeds caused by the standing stubble, resulted in reduced evaporation on the stubble plot compared with the fallow plot. Roughness parameter and zero plane displacement for 50 cm high stubble are typically in the range of 6 and 31 cm and for a fallow surface 0.4 and 0 cm respectively (see Oke, 1978). The benefits of maintaining upright crop residue to improve soil moisture conservation are well documented (Greb *et al.*, 1967; Masse and Cary, 1978).

The availability of soil water was another factor affecting the evaporation rates from the two surfaces. In the top 60 cm of the soil profile there was, on average, 8% more water (v/v) on the fallow plot than on the stubble plot (Table 2). The low rate of water loss from both plots during the 3 days of chinook conditions indicated that evaporation during the period was mainly confined to the falling rate and diffusion stages (Ritchie, 1972) and thus evaporation was for the most part not energy-limited. Most of the absorbed energy at the surfaces was dissipated as soil heat flow and sensible heat. Evaporation rates during this period were, therefore, limited mostly by the effects of surface cover and the availability of soil water for evaporation.

As evaporation dissipated little of the absorbed energy at the surface of the plots, soil surface temperatures reflected changes in net radiation. Thus, the greater amount of net radiation on the fallow surface, compared with the stubble, resulted in higher soil temperatures and greater soil heat flow in the fallow plot (Figure 3a, 3b). Soil temperatures were most variable at the surface with a reduced amplitude of the temperature wave, and increased thermal lag with increasing

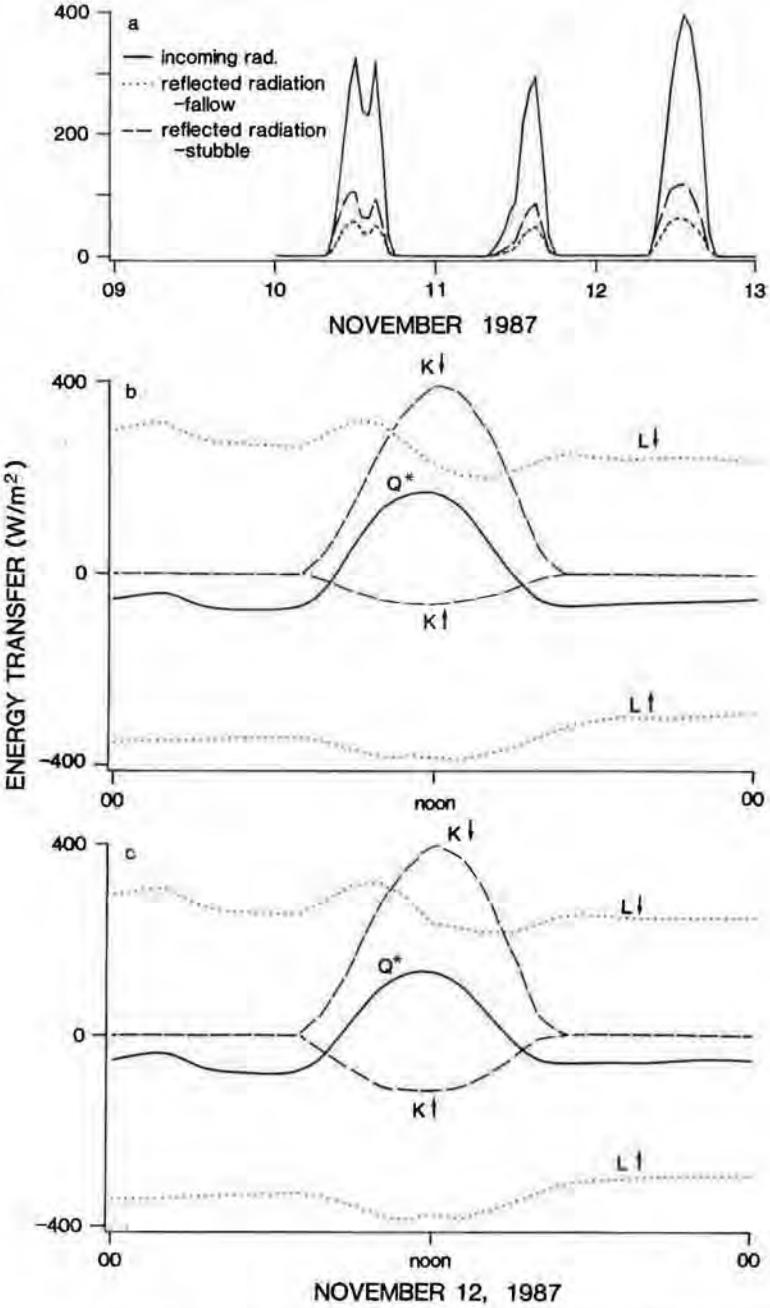

FIGURE 2. (a) Mean hourly incoming and reflected solar radiation (W/m²) for fallow and stubble plots, November 10–13.

(b) Radiation balance for the fallow plot, November 12 1987. K↓ is incoming solar radiation, K↑ is reflected solar radiation, Q* is all-wave net radiation, L↓is incoming long-wave radiation, L↑ is emitted long-wave radiation.

(c) Radiation balance for the stubble plot, November 12 1987.

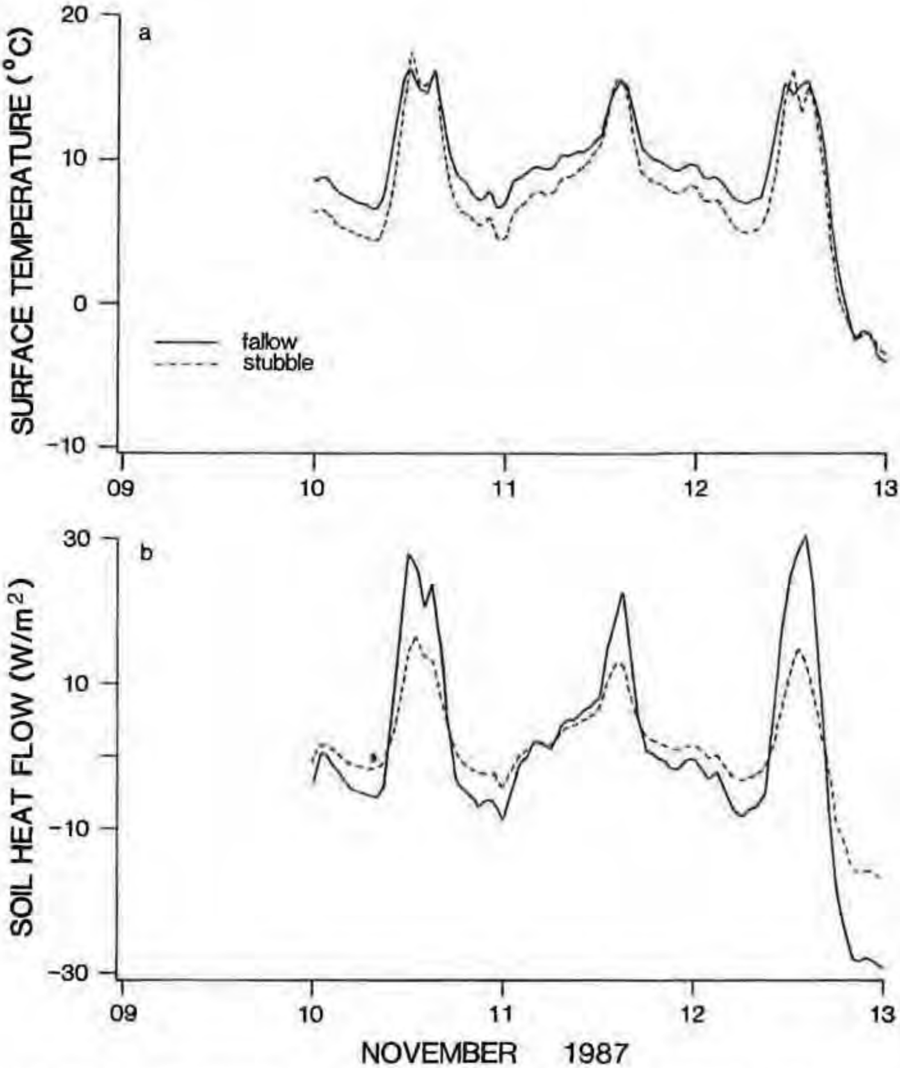

FIGURE 3. (a) Mean hourly soil surface temperatures (°C) and (b) mean hourly soil heat flow (W/m^2) for fallow and stubble plots, November 10–13 1987.

depth. Soil surface temperatures on the stubble plot exceeded those of the fallow plot only during the midday period when incoming solar radiation was at a maximum and surface shading by the standing stubble was minimized. As has been noted elsewhere (Waggoner *et al.*, 1960) a straw or crop-residue mulch reduced evaporation, increased sensible heat generated at the surface, and reduced the amount of heat penetration into the soil. The temperature of the mulch was high but led to a cooling effect on the soil below. The crop residue absorbed radiation but transmitted little.

Temperature profiles for the fallow and stubble indicated that during the midday hours the soil surface temperatures under the stubble and residue were similar to those of the fallow plot, probably because of the high temperature of the surface residue. The lack of turbulent transfer of heat within the litter or crop residue allowed heat to build up at this time. However, the soil temperatures to a depth of 50 cm were cooler under stubble at all times of the day. Surface temperatures seldom varied more than a few degrees from air temperatures owing to rapid dissipation of net radiation via soil heat flow and especially sensible heat during times of high windspeeds. As noted elsewhere (Oke, 1978), diurnal variation in soil heat flux was greater under the bare soil conditions (Figure 3b).

b) *November 16 to 22 1987*

A 6.7 cm snowfall (5.4 mm water equivalent) recorded on November 16 1987 at the nearby (1 km) Agriculture Canada Weather Observing Site resulted in 2 to 5 cm of snow accumulation on the fallow plot whereas snow depth on the stubble plot measured 20 to 25 cm. The strong wind accompanying and immediately following the snowfall allowed the stubble plot to trap much of the blowing snow. Late in the day of November 17 and early in the morning of November 18, the presence of a chinook arch along with increasing westerly windspeeds, air temperatures and vapor pressure deficits signalled the onset of chinook conditions. During the next 4 days air temperatures never fell below 0°C, vapor pressure deficits of greater than 1 kPa were observed, and windspeeds recorded at a height of 1.5 m were usually greater than 5 m/s and hourly averages of greater than 10 m/s occurred in the early morning hours of November 18.

During the 2 days following the November 16 snowfall much of the snow was removed from the fallow plot by melting, evaporation, sublimation, and wind erosion. Figure 4 displays incoming and reflected solar radiation from the two plots for the days under consideration. On the first day following the snowfall, November 17, the plots had similar reflected solar radiation (a value of 80% albedo was calculated for midday) indicating snow cover on both surfaces. However, the following day (November 18), reflected radiation from the fallow plot was significantly reduced (18% albedo) as snow was removed from the plot to expose the dry soil surface below. Weighing lysimeter data (Figure 1b) indicated that after the snow was removed from the fallow plot by midday on November 18, only 1.5 mm water was lost from the lysimeter during the next 3 days. It was not until 3 days after the fallow plot was snow-free that the stubble plot was also free of snow and the values of reflected solar radiation approached those of the fallow plot (Figure 4). Even when the snow cover was completely gone from the stubble plot the albedo was still slightly higher than that of the fallow plot owing to the inherent difference in reflectivity of the two surfaces.

The microclimate of the snow surface on the stubble plot was complicated by penetration of solar radiation into the snowpack, the constantly

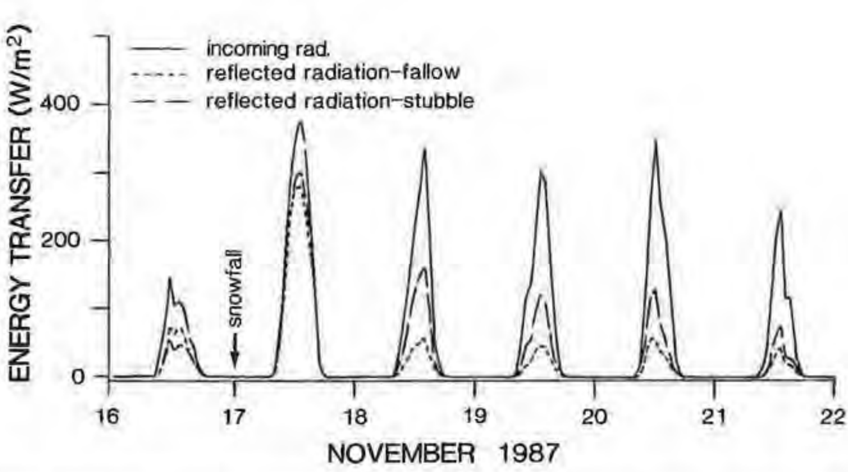

FIGURE 4. Mean hourly incoming and reflected solar radiation (W/m²) for fallow and stubble plots, November 16–22 1987.

changing height and nature of the active surface, and internal water movement and phase changes. Percolation of meltwater may cause water movement inside a snowpack. Energy uptake or release results from phase changes of water within the pack (e.g., freezing, melting, sublimation, evaporation or condensation).

On the clear cold day following the snowfall, November 17, the effect of the high albedo on net radiation of the stubble plot was as displayed in Figure 5a. The surface had a high albedo of more than 95% for the morning and afternoon hours with a midday minimum of only 80%. This resulted in very little of the incident solar radiation being absorbed by the snowpack. Generally, the daytime net radiation surplus of snow-covered surfaces is small by comparison with most other natural surfaces. Here, however, there was actually a deficit in net radiation. Most of the incident solar radiation was reflected and the snowpack emitted more long-wave energy to the clear sky than it received (Monteith, 1973). The source of the energy was the stored energy in the snowpack (Hicks and Martin, 1972; McKay and Thurtell, 1978), which can be considerable, especially when the snow falls under mild conditions such as those that existed on November 16. The result (Figure 5a) was a negative net radiation, or energy loss, from the snowpack during all hours of the day on November 17. Only during the midday period, when incident solar radiation was greatest and albedo of the surface least, did the energy loss approach zero. Thus, the stored energy in the snowpack was reduced by radiative losses as well as sensible and latent heat losses. The result of the loss of energy from the snowpack was to increase the "cold content" of the snowpack (O'Neill, 1972), the "cold content" being the temperature that is attained by the snowpack during a period of net energy loss.

During the active melting of the snowpack on the stubble plot

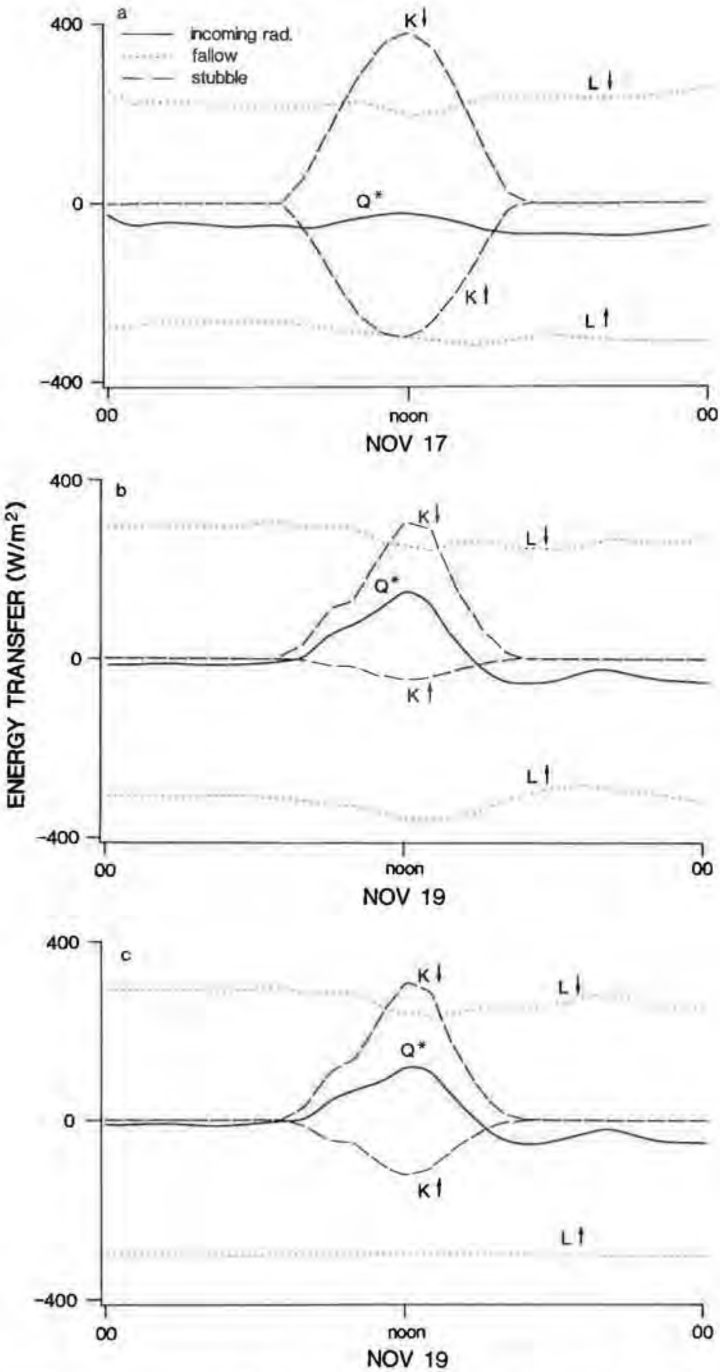

FIGURE 5. Radiation balance where K↓ is incoming solar radiation, K↑ is reflected solar radiation, Q* is all-wave net radiation, L↓is incoming long-wave radiation, and L↑ is emitted long-wave radiation. (a) stubble plot on November 17, (b) fallow plot on November 19, (c) stubble plot on November 19, 1987.

(November 18, 19, and 20), when air temperatures were always 0°C or greater, the weighing lysimeter data (Figure 1b) did not indicate a steady loss in weight as the snow sublimated or melted and then evaporated. Indeed, there were times when the lysimeters appeared to gain weight. The surface vapor pressure of the melting snow surface was the saturation value of 0°C or 611 Pa. On several occasions the warm air above the snow surface had a greater vapor pressure than the wet snow surface. Thus an air-to-snow vapor pressure gradient existed and resulted in a downward flux of moisture and hence condensation on the surface. In absolute terms, the amount of water condensed out of the atmosphere during these times would have been a function of near-surface turbulence and the magnitude of the vapor pressure gradient. As the latent heat of vaporization (2.501×10^6 J/kg) released upon condensation is approximately 7.5 times larger than the heat of fusion (3.33×10^5 J/kg) needed for melting snow, condensation during these periods not only added water to the snowpack but also provided a significant energy source and thus hastened the melting process.

It is interesting to note that during the days of active melting of the snowpack on the stubble plot the atmosphere was a source of water vapor for the melting snowpack while simultaneously being a sink for water vapor from the adjacent warm fallow plot. During a 24-hour period, Munro (1975) in a study of a melting glacier indicated that the convective transfer of energy, sensible heat, and latent heat provided 29% of the total energy used to melt the snow with the remaining 71% coming from net radiation absorpotion. These data of the ranking of heat sources during the melting agree well with the review by Paterson (1969) of 32 glacier energy balance studies. Oke (1978) has suggested that such heat sources are likely to apply to snow melt over other open surfaces such as prairie sites.

Although it was not possible to distinguish the effects of sublimation, melting, and evaporation or infiltration from the data presented here for the stubble plots, it was clear that more moisture was added to the soil in the stubble plot than in the fallow. Some of the weight increases in the stubble plot were the result of deposition of soil by the strong winds of November 18. The airborne soil particles originated from large cultivated fields east of Lethbridge where serious soil erosion problems exist. Little soil loss was observed on the adjacent fallow plot as its size conformed to recommendations for strip-farming practices (see above). The darkened soil surface resulted in a reduced value of reflected solar radiation on November 18 for the snow-covered stubble plot (Figure 4). Variation in albedo was also due to physical changes in the state of the snow surface when there was surface melting. Even a thin film of meltwater on the snow surface serves to reduce albedo to a value closer to that of water (0.01 to 0.10 for small zenith angles) (Oke, 1978).

Complete energy balances for the fallow and the stubble plots on November 19 (Figure 5b, 5c) show that incoming solar radiation reached a midday maximum of 307 W/m². Long-wave radiation emitted from the surface of the snow-covered stubble plot was constant at 309.5 W/m² based on a surface temperature of the melting snowpack of 0°C and an emissivity of 0.9 (Oke, 1978), whereas the long-wave radiation emitted from the bare fallow plot varied as a function of the measured soil surface temperature of the plot from a minimum of 293 W/m^2 to a maximum afternoon value of 375 W/m^2. The largest changes in the surface energy balances resulted from differences in surface albedo. The albedo of the snowpack on the stubble plot averaged 39% over the course of the day while the albedo of the dry soil surface of the fallow plot averaged only 15%. The large

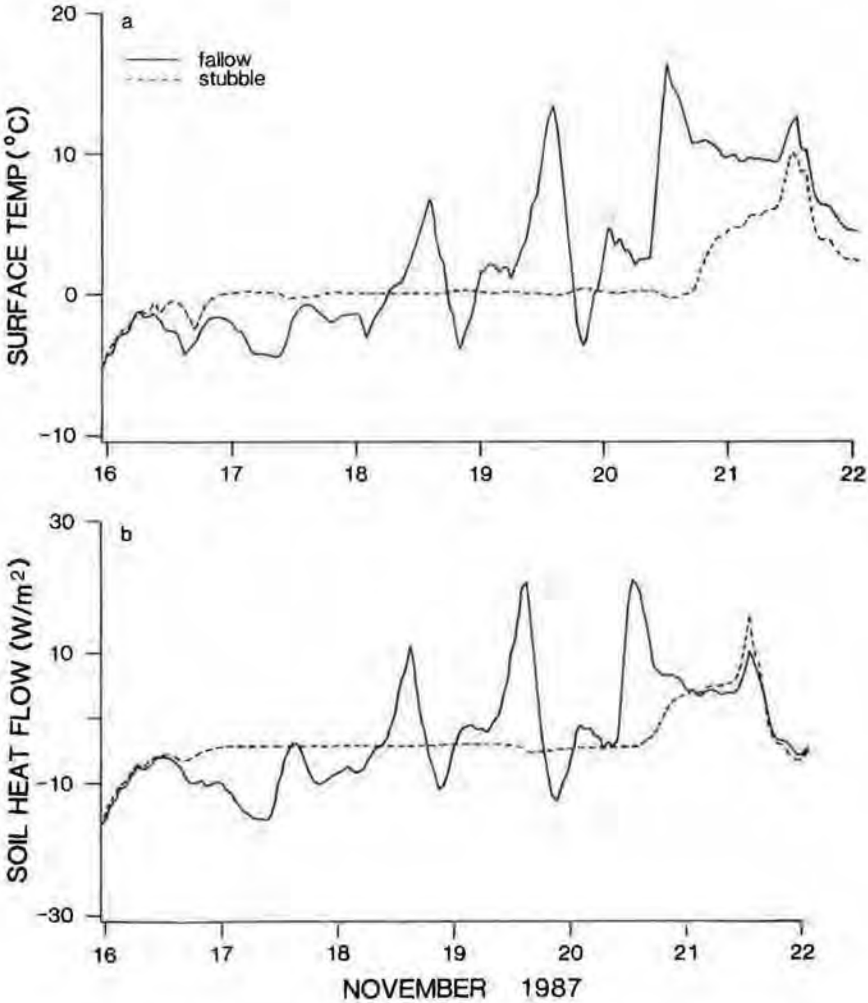

FIGURE 6. (a) Mean hourly soil surface temperatures (°C) and (b) mean hourly soil heat flow (W/m^2) for fallow and stubble plots, November 16–22 1987.

difference in surface albedos resulted in differences in absorbed energy. The total daily absorbed energy for the fallow surface was 604.4 kJ/m^2, whereas 475.8 kJ/m^2 was observed on the stubble plot.

The decreased albedo of the fallow plot on November 18 (Figure 4) resulted in increased absorption of energy at the soil surface. This was reflected in soil temperature and soil heat flux data (Figure 6). Snowfall on November 16 had the effect of damping the surface temperature fluctuations. The difference in snow accumulation on the two plots accounted for the differences in thermal damping between them. The very low thermal conductivity and diffusivity of snow (especially when fresh) makes it an effective insulator for the ground beneath (Oke, 1978). This was especially true at night when radiative exchange was concentrated in the surface layer of the snow. Under fallow conditions and only a few centimeters of snow, a soil surface temperature of -5°C was recorded in the early morning hours of November 17 in response to cold air temperatures (-16°C). However, under stubble conditions and 20 cm of snow, the soil surface temperatures remained at 0°C for several days.

As a result of the differential snow accumulation, net soil heat flow was uniform and small in magnitude (Figure 6b) under stubble and directed upwards from the warm soil beneath. Under fallow conditions, the soil heat flux quickly resumed the typical diurnal pattern of positive (down) during the daylight hours and negative (up) at night. Only when all snow was melted from the stubble plot on November 20 did soil heat flux and near-surface soil temperatures resume a diurnal pattern in response to variation in net radiation.

The changes in snow cover and hence absorbed energy at the surface were expressed in temperature profiles. For example, during the midday period of November 18 and 19 when net radiation was greatest, soil surface temperatures of the fallow plot were often 5 to 7°C warmer than those of the stubble surface. Although the soil surface temperature under the snowpack on the stubble plot varied little from 0°C during the course of the day, the diurnal variation of surface temperature on the fallow plot was greater than 7°C.

The net effect of the snowfall and the subsequent chinook was to increase the soil moisture in the surface layers of the soil in the stubble plot over that in the fallow plot (Table 3). Although there was a slight increase in soil moisture in the top 30 cm of the fallow plot (16.00 to 16.27% v/v) (Tables 2 and 3),

TABLE 3. Percent soil moisture (v/v) from neutron attenuation for the stubble and fallow study plots, November 25, 1987 (means of measurements taken at 5 locations within each plot).

	Fallow		Stubble	
Depth (cm)	Mean	S.E.	Mean	S.E.
0–30	16.27	0.97	13.20	2.00
30–60	19.80	0.30	11.00	1.13
60–130	16.10	0.60	11.87	0.73

precipitation that fell on this plot was removed by wind, sublimation, and evaporation. While much of the snow cover on the stubble plot was also removed by evaporation and sublimation, the increase of 3.4% in soil moisture in the top 30 cm added to soil moisture reserves.

It should be noted at this point that the increased soil moisture resulting from snowfall and subsequent chinook conditions was strongly influenced by the condition of the soil. During the observation period discussed here, the soil profile was not frozen. A frozen soil will limit the infiltration of meltwater (Gray *et al.*, 1985; Nicholaichuk, 1986) and result in greater water losses through runoff and evaporation. Additional study is required to assess the effects of surface cover and chinooks on soil moisture during times of frozen soil. On average, however, the number of days that a profile is frozen in the Lethbridge area is small (90 days/year, Grace and Hobbs, 1986). Thus, the maintenance of stubble may be of benefit in trapping snow and increasing overwinter soil moisture reserves during times of an unfrozen soil profile in the chinook-dominated, semi-arid environment of southern Alberta.

4. SUMMARY

The two surfaces of stubble and fallow showed slight changes in near-surface microclimates in response to chinook conditions when the soil surface was dry. Most notable were changes in net radiation in response to the differences in albedo of the two surfaces and the proportioning of dissipated energy in soil heat flow and sensible heat as there were only slight changes in soil water and hence evaporation. The difference in evaporation rates was attributed to availability of soil water and to the effects of the surface cover. The chinook conditions did not seriously affect soil moisture status. However, the near-surface microclimate of the two surfaces was significantly different in response to chinook conditions following precipitation of snow.

The most dramatic changes were a result of the snowfall accumulation on the respective surfaces and the subsequent disposition of energy and moisture. The minimal accumulation of snow on a fallow surface resulted in rapid removal of the snowcover with little of the moisture being added to the soil profile. The ability of the stubble surface to accumulate significantly more snowfall than the fallow surface required that more energy be spent in the subsequent evaporation, sublimation, melting, and infiltration. It is clear from the microclimatic observations presented that the ablation of snowpack under chinook conditions occurs through a complex and varied series of physical processes. Dry, cool, and windy periods provided conditions for direct sublimation whereas other warmer periods allowed vapor flux from the atmosphere to the snowpack. Absorption of net radiation into the snowpack also provided energy for melting. In addition, penetration of a portion of the incoming solar radiation flux through the snowpack and absorption by the surface below the snowpack undoubtedly combined with the warm soil and negative heat flow directed towards the soil surface to cause some melting from the bottom of the snowpack. Although much of the potential moisture was lost to the atmosphere during chinook conditions, infiltration of *in situ* meltwater did provide a significant contribution to soil moisture reserves of the unfrozen soil profile.

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Integration of Automated Station Data Into Objective Mapping of Temperatures for an Arctic Region

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ABSTRACT

Mapping and modelling of the surface climate in relation to features of bioclimatic and hydrologic interest in the Arctic is difficult because of the relatively low density of meteorological stations. Automated climate stations ("autostations") provide a useful complement to the permanent network. Summer temperature data from two automated stations in the interior of Baffin Island have been used to test the local and mesoscale validity of the existing regional climatology. Objective analysis was used to produce maps of the temperature fields with and without the autostations and to evaluate the existing subjective maps of temperature normals. The results demonstrate the usefulness of the approach in resolving mesoscale detail and regional spatial trends and point to a generalized approach to mesoscale modelling.

RÉSUMÉ

La cartographie et la modélisation des climats à la surface combinés aux facteurs bioclimatiques et hydrologiques sont très difficiles dans l'Arctique à cause de la rareté des stations d'observation météorologique. Toutefois, les stations d'observation météorologique automatiques présentent un heureux complément au réseau déjà existant. Les données de température estivale de deux stations automatiques situées sur l'Ile de Baffin ont été analysées afin de vérifier certaines hypothèses climatiques régionales à l'échelle locale et à l'échelle moyenne. Des analyses objectives ont permis de produire une série de cartes de températures provenant des stations automatiques et des données conventionnelles, et d'évaluer la subjectivité des cartes de températures normales déjà existantes. Les résultats démontrent l'utilité de cette approche en ce qui concerne la résolution des détails à l'échelle moyenne et des tendances spatiales régionales, et fournissent la base d'une approche généralisée quant à la modélisation à l'échelle moyenne.

INTRODUCTION

In the Canadian Arctic, we are faced with the problem of trying to describe climate adequately from a relatively short series of observations at too few stations. While the records are now sufficiently long to permit regional-scale descriptions and mapping of at least the main climatic elements, the density of meteorological stations is usually insufficient to permit detailed analysis and mapping. For many applications, the scale of interest is the *mesoscale*, a term used here loosely to mean the spatial scale of order 10^1 to 10^2 km. Many significant features of the arctic landscape occur on this scale, particularly near its lower end. Examples are glaciers and ecological "oases", which may have or reflect an influence that extends far beyond their boundaries.

Recognition of the complexity of landscape in the Arctic coupled with an understanding of micro- and mesoscale processes leads to the conclusion that the regional maps are only a first approximation to climate. A mesoscale grid superimposed on such maps and resolved cell by cell will almost certainly result in a description of climate that differs on the local level from simple interpolation of the regional scheme. Integrated over the whole grid, such an approach will likely alter the regional picture as well.

This study is concerned with finding methods to reveal the detail within the mesoscale grid. The conceptual framework is one that integrates data from the long-term operational network with short-term observations from temporary stations at key locations. Objective interpolation methods and the possible incorporation of point climate models permit the analysis to be extended over the region. This paper reports on the first phase of the study, an evaluation of daily mean temperatures during the summer period.

FIELD INVESTIGATIONS

The eastern Canadian Arctic, comprising Baffin Island and adjacent waters (Figure 1), spans about 15 degrees of latitude and includes a broad range of climatic zones (Maxwell, 1980, 1981). Baffin Island, with an area of nearly 500,000 km^2, contains several distinct physiographic units, from the glaciated highlands of the east coast and central plateau, to the lowlands of the west-central region. At least three broad bioclimatic zones are represented within the island (Energy, Mines and Resources, 1974). The region presents an increasingly well-documented record of environmental change (Andrews, 1985) and is a logical focus for ongoing monitoring of such changes.

Attention was drawn initially in this study to an apparent mesoclimatic anomaly in the south-central interior region, including Nettilling Lake (Figure 1). Field work there in 1985–86 revealed some representative low arctic plant species well beyond their previously recorded limits. Short-term measurements confirmed the existence of a locally warm summer climate, with daily temperatures in July averaging about 2°C above regional means. Energy budget calculations indicated

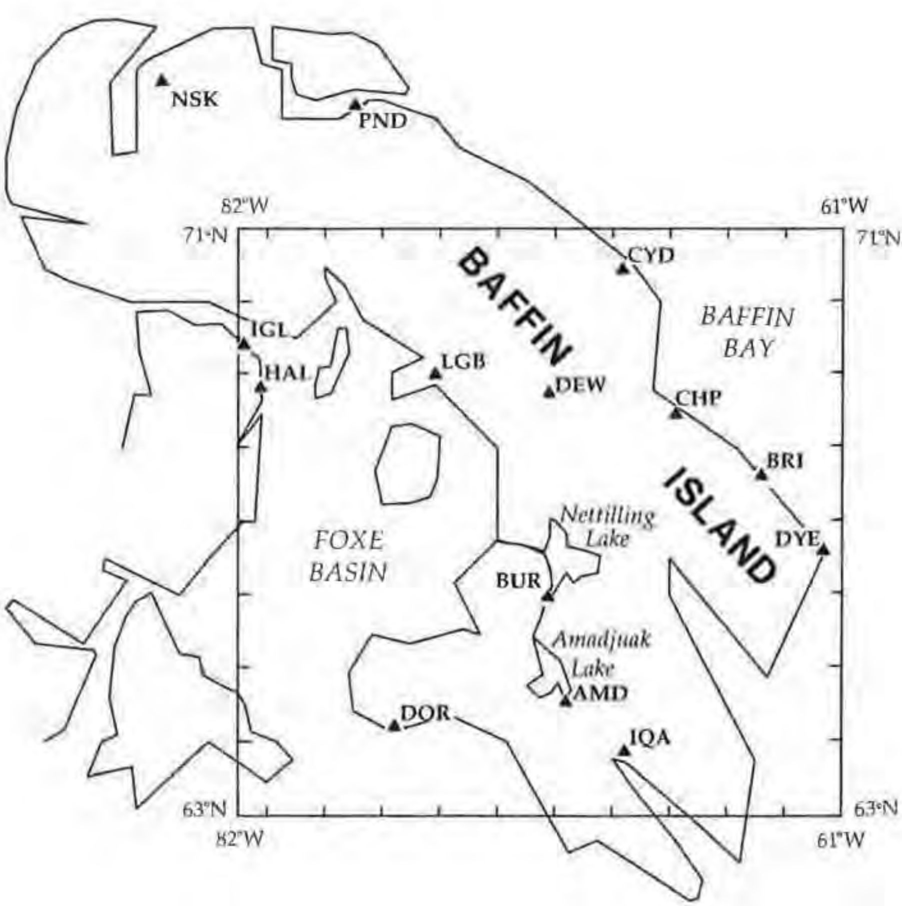

FIGURE 1. The Baffin Island Region, showing meteorological stations. Abbreviations are as given in Table 1.

that heat advection in the numerous streams and the large heat capacity of the lakes contribute substantially to the moderation of climate in the Amadjuak Lake – Nettilling Lake drainage basin (Jacobs and Grondin, 1988).

Automated climate stations ("autostations") were set up near Amadjuak Lake and on the south end of Nettilling Lake at Burwash Bay in July of 1987. These stations record air and soil temperatures, relative humidity, global solar radiation, wind speed and direction, and precipitation at 3-hourly intervals. At the Amadjuak Lake site (Figure 2), nearshore lake temperatures and snow depth are measured in addition to the other variables. A third autostation was installed in 1989 near the northwest margin of the Barnes Ice Cap.

Temperature measurements at the autostations differ from those at the permanent stations in that the standard liquid-in-glass thermometer in a Stevenson screen is replaced by a thermistor in a Gill multi-plate radiation shield. Comparative observations elsewhere have revealed differences in temperatures measured with different sensors and shields, suggesting a conservative limit of resolution of 1.0°C for unaspirated shields (McTaggart-Cowan and McKay, 1976). Baker and Ruschy (1989) reported that mean daily temperatures measured with a thermistor in a radiation shield averaged about 0.1°C above those from an unaspirated thermometer in a Stevenson screen. For the shields on the autostations in the present study, the manufacturer's claim of a maximum departure from ambient of less than 0.7°C under light winds and strong solar heating was confirmed by spot readings in the field with a sling psychrometer.

FIGURE 2. The automated climate station at Amadjuak Lake.

Data were retrieved from the Amadjuak Lake and Nettilling Lake stations in the summer of 1988. Failure of the datalogging system at the former meant that only 32 days of data were obtained. Some interruptions occurred in the Nettilling Lake station record as well, but a nearly complete year's record of air temperature, humidity, and wind was recovered.

ANALYSIS AND RESULTS

The period chosen for initial study was 19 July – 19 August 1987, when the Amadjuak Lake autostation was still functioning. This corresponds well with "summer" in the region. Records were available for purposes of comparison from the 12 permanent meteorological stations shown in Figure 1. These include the tier of 6 DEW-line sites from Hall Beach to Cape Dyer, as well as Igloolik, Nanisivik, Pond Inlet, Clyde, Iqaluit, and Cape Dorset (Table 1).

Pairwise correlations of daily mean temperatures for the 32 days of record were carried out among the 14 stations (12 permanent and 2 autostations), constituting 91 different pairs. Table 2 shows the results of the correlations. For $n - 2 = 30$ degrees of freedom, the critical value of r at the 1 percent level is 0.449. This value was exceeded in 23 cases.

TABLE 1. Climatological stations referred to in this study.

Station	Latitude °	Latitude '	Longitude °	Longitude '	Elevation m.a.s.l.
Amadjuak Lake[a](AMD)[b]	64	34	70	29	120
Broughton Island (BRI)	67	33	63	47	598
Burwash Bay[a](BUR)	65	57	71	18	47
Cape Dorset (DOR)	64	13	76	32	46
Cape Dyer (DYE)	66	35	61	37	393
Cape Hooper (CHP)	68	26	66	47	401
Clyde (CYD)	70	28	68	37	25
Dewar Lakes (DEW)	68	39	71	14	518
Hall Beach (HAL)	68	47	81	15	8
Igloolik (IGL)	69	23	81	48	21
Iqaluit (IQA)	63	45	68	33	34
Longstaff Bluff (LGB)	68	57	75	18	162
Nanisivik (NSK)	72	59	84	38	639
Pond Inlet (PND)	72	41	77	59	4

[a] Autostation

[b] Abbreviation for purposes of Table 2.

TABLE 2. Interstation correlation for daily mean temperatures, 19 July – 19 August 1987.

	AMD	BUR	CYD	DOR	DEW	DYE	BRI	CHP	HAL	IGL	IQA	LGB	NSK
AMD	1												
BUR	0.74[a]	1											
CYD	0.39	0.23	1										
DOR	0.18	0.20	0.04	1									
DEW	0.58[a]	0.73[a]	0.18	0.26	1								
DYE	0.34	0.38	0.52[a]	0.09	0.41	1							
BRI	0.54[a]	0.53[a]	0.41	0.01	0.63[a]	0.77[a]	1						
CHP	0.51[a]	0.29	0.66[a]	0.16	0.30	0.68[a]	0.69[a]	1					
HAL	0.21	0.16	0.06	0.64[a]	0.33	0.18	0.01	0.19	1				
IGL	0.15	0.18	0.31	0.55[a]	0.36	0.31	0.11	0.44	0.90[a]	1			
IQA	0.58[a]	0.20	0.19	0.36	0.10	0.03	0.08	0.33	0.35	0.23	1		
LGB	0.26	0.57[a]	0.22	0.26	0.73[a]	0.01	0.14	0.32	0.44	0.67[a]	0.05	1	
NSK	0.13	0.21	0.15	0.04	0.46[a]	0.01	0.01	0.17	0.10	0.09	0.34	0.58[a]	1
PND	0.02	0.18	0.02	0.23	0.37	0.26	0.17	0.10	0.52[a]	0.38	0.27	0.20	0.54[a]

[a] Significant at the 1% level or better.

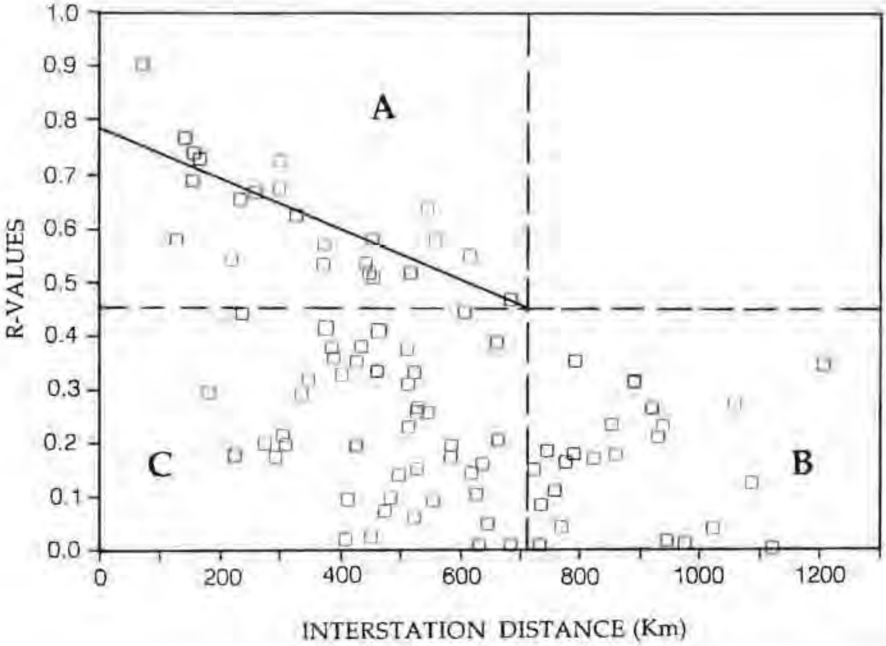

FIGURE 3. Correlation - distance plot for 14 Baffin Region stations for mean temperature for the 1987 summer period (19 July – 19 August).

In Figure 3 the correlation coefficients are plotted against interstation distances. Group "A" are the pairs which correlated significantly and conform to a linear correlation-distance function represented by the solid line, a least-squares fit to those points. The point where that line falls to $r = 0.449$ defines the limiting distance (about 710 km) beyond which station pairs (Group "B") did not correlate significantly. Some nearer pairs (Group "C") are also in poor agreement, presumably due to factors other than interstation distance such as elevation, distance from the coast, or local influences.

The significant correlations were used to delineate a subregion centred approximately on the two autostations, and enclosed by the box in Figure 1. Maps of the temperature fields were constructed using the technique of optimum interpolation with kriging (Delhomme, 1978; Fortin, *et al.*, 1983), which has the advantage of giving the best possible representation for each station. Contours were smoothed using a cubic spline procedure. Both procedures are part of a commercial mapping software package called Surfer.[1]

1 "Surfer" is a PC-based mapping software system produced by Golden Software, Inc., P.O. Box 281, Golden, Colorado 80402, USA.

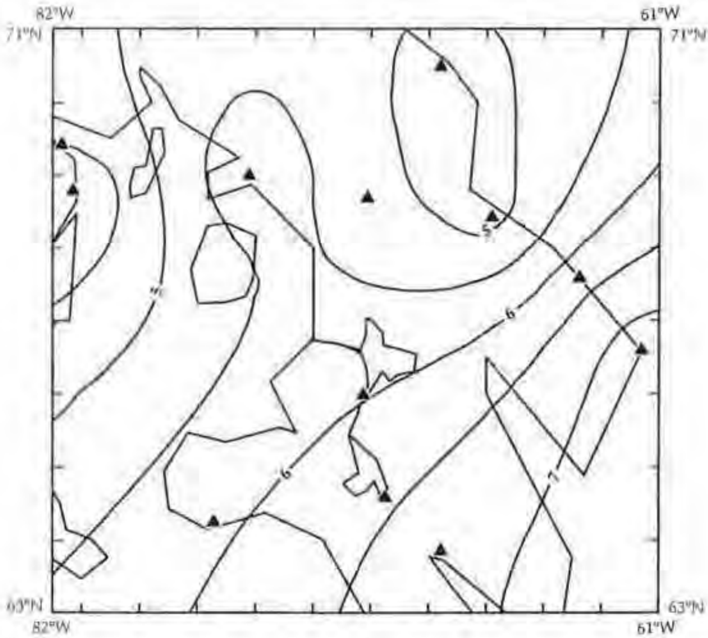

FIGURE 4. Mean daily temperature (°C) for the 1987 summer without autostation data.

Figure 4 shows the resulting temperature pattern based on the permanent station network without autostations. The effect of adding the autostations is shown in Figure 5. Similar patterns (not shown) resulted when the autostations were incorporated singly. Figure 6, a mapping of the difference between the fields portrayed with and without the autostations, suggests the extent and magnitude of the temperature anomaly.

A stepwise multiple regression was used between the summer temperature data for each autostation (dependent variable) and those of the permanent stations (independent variables) to see if the latter might be used as predictors of temperature at the autostation sites. It was found that temperatures at Burwash Bay were predicted within the 95 percent confidence limit by a single station, Dewar Lakes. The resulting equation is:

$$T_{BUR} = 6.51 + 0.56T_{DEW}$$

For Amadjuak Lake, both Dewar Lakes and Iqaluit contribute. The equation is:

$$T_{AMD} = 3.70 + 0.33T_{DEW} + 0.43T_{IQA}$$

Scatterplots of predicted versus observed temperatures (not shown) gave a nearly even distribution in both cases.

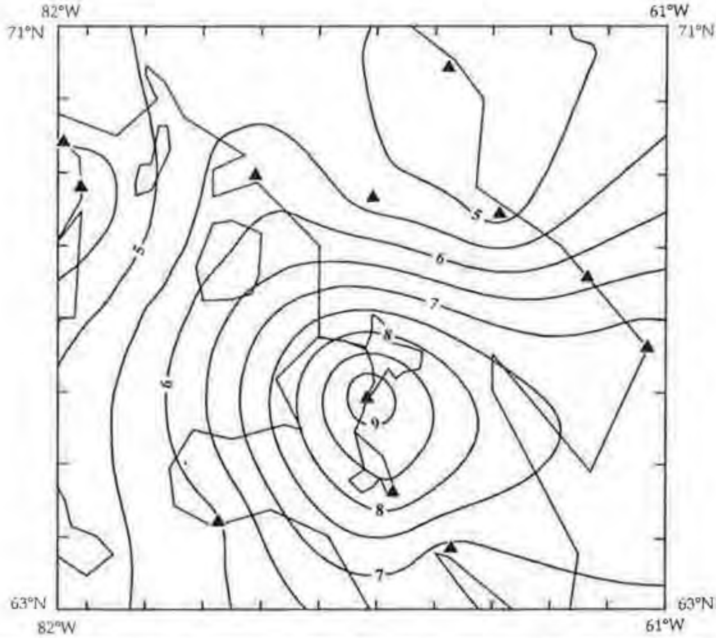

FIGURE 5. Mean daily temperature (°C) for the 1987 summer including the Amadjuak Lake and Burwash Bay autostations.

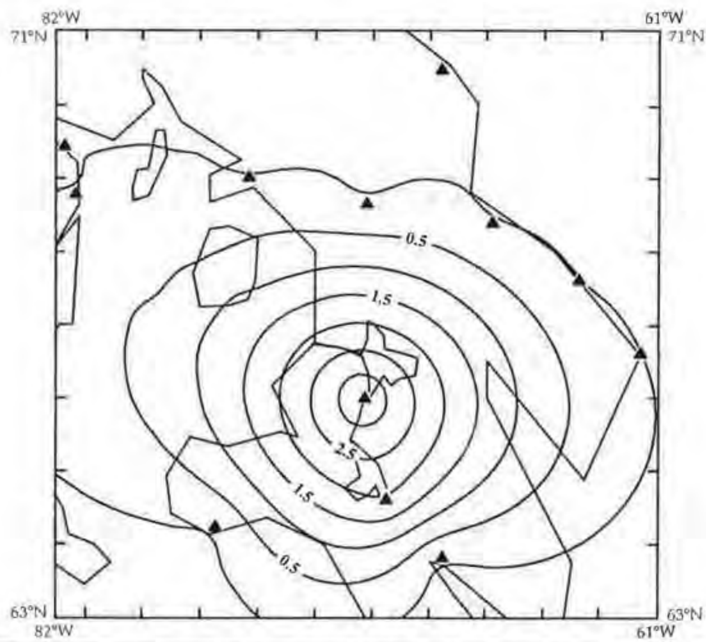

FIGURE 6. Mapping of differences (°C) between plots with (Figure 5) and without (Figure 4) autostation data.

It was noted that, although the technique of temperature measurement differs between permanent stations and autostations, the results are comparable. Therefore the magnitudes of the point anomalies represented by the Amadjuak Lake and Nettilling Lake records must be real. Interpolation incorporating those anomalies results in a spatial pattern (Figure 5) that is superior to the one without them (Figure 4) but which undoubtedly still simplifies the true situation. It is recognized as well that observations from only one year have limited value for climatological purposes, and several years of observations will be required before the autostations can be replaced by the "virtual" stations implicit in the regression equations.

In order to assess the effectiveness of optimum interpolation for long-term means, the computer mapping procedure was applied to July mean temperatures (1951–80 normals) for the permanent stations to produce an objective map (Figure 7). This can be compared with a map published by Maxwell (1980) for approximately the same period but based largely on subjective analysis (Figure 8).

The objective mapping reveals several features of the temperature field not shown in the subjective map, including a general latitudinal gradient. The

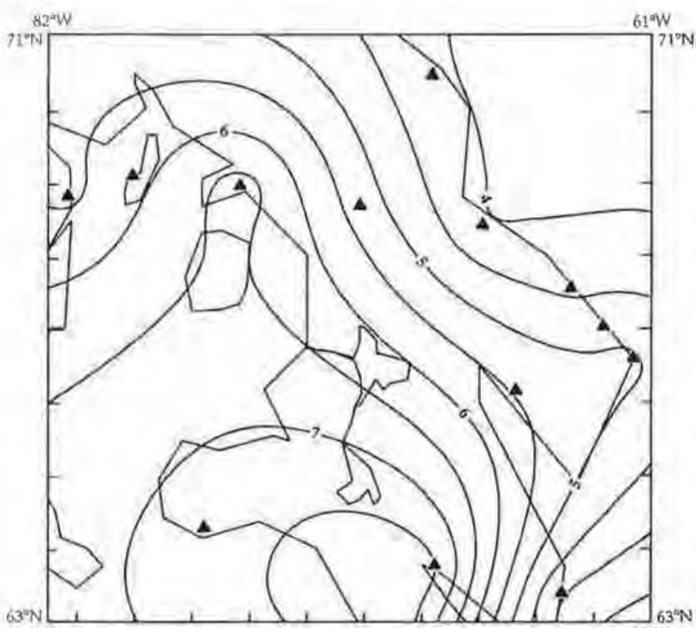

FIGURE 7. Objective mapping of mean daily temperature (°C) for July from Atmospheric Environment Service 1951–1980 station normals.

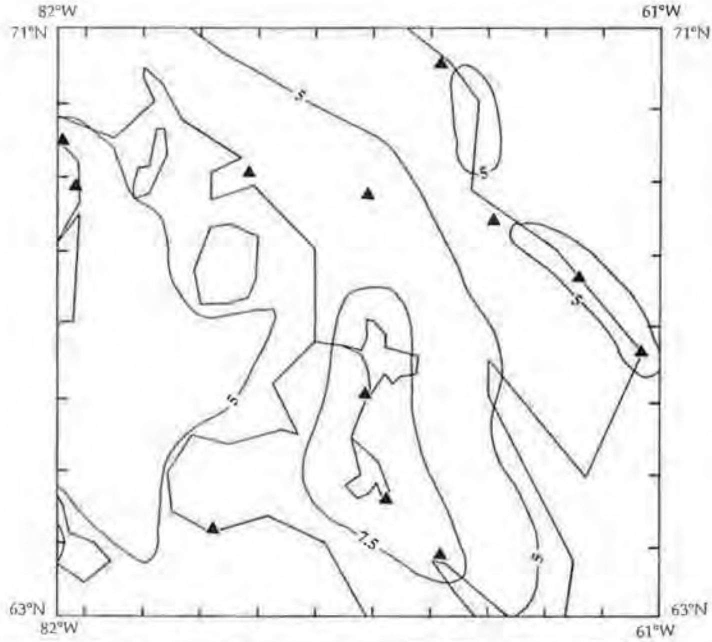

FIGURE 8. Subjective mapping of mean daily temperature (°C) for July after Maxwell (1980).

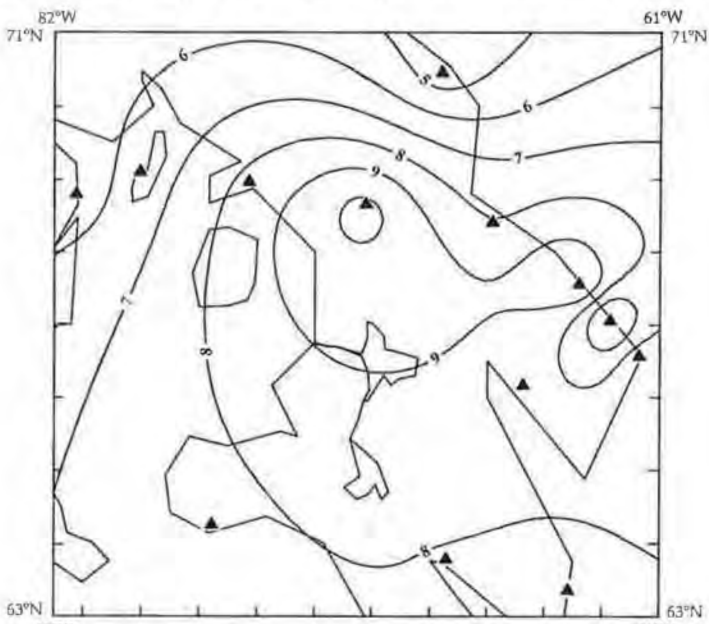

FIGURE 9. Objective mapping of mean daily potential temperature (°C) for July for the period 1951–1980.

cold influence of Baffin Bay and Davis Strait on the east coast is more pronounced in the objective map, but a similar influence from Foxe Basin was not picked up at all. The suggestion of a positive anomaly in the interior lowlands, shown in Maxwell's map, was not apparent in the objective map because no data from that area were actually included in the set of station normals. In this case, the 1987 autostation data presented in Figures 5 and 6 are confirmation of the assumptions underlying the subjective mapping.

In recognition of the considerable range of station elevations in the region (Table 1), objective mapping of potential temperatures for the 1951–80 means was also carried out. In potential temperature calculations the temperature for each station is standardized by bringing it down from station elevation to sea-level at the dry adiabatic rate. The resulting map (Figure 9) reveals large positive anomalies at the higher central and east coast stations that are a reflection of topoclimatic influences and the regional inversion.

The effects of local site factors, implicit in the correlation-distance plots (Figure 3), point to the need for further investigations of terrain characteristics (elevation, slope, surface cover) and their inclusion in the climate models. Correlative studies of vegetation, based on field surveys expanded with remote sensing data, could assist greatly in determining the spatial variability of local and mesoscale climate.

Temperature is only one of the elements of climate. Mapping of precipitation, winds, and radiation, while amenable to an objective approach, may require different assumptions. It was found for example in an earlier study (Jacobs, 1989) that there is no significant spatial pattern in the data for seasonal snowfall and depth of snowcover from the stations of the Baffin Region. This points to a serious deficiency in the climatic descriptions of the region and calls into question certain conclusions about wildlife habitat, regional water balance, and related subjects.

CONCLUSION

These results demonstrate that the accuracy and spatial resolution of climatic maps and models for the Arctic can be significantly improved through objective analysis incorporating data from autostations. Further refinement can be made through the use of additional autostations at carefully chosen sites, along with ancillary studies of mesoscale terrain and vegetation factors and their incorporation in the analysis. There will of course always be a need for some degree of subjectivity, as the network of stations can never be dense enough to be truly representative of the local climatic diversity of the region.

ACKNOWLEDGEMENTS

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Temporal Relationships Among Multiple Drought Types During an Average Drought

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ABSTRACT

The primary purpose of this paper is to examine the concurrent temporal patterns of drought subtype severity during the first six months of an average drought event. The methodology employed in this study is designed to help search for the underlying signals or patterns across a sample of drought events. Monthly values of the Palmer Moisture Anomaly Index (ZINX), Palmer Drought Severity Index (PDSI), Palmer Hydrologic Drought Index (PHDI), and standardized scores (z-scores) of precipitation (PREZ) and temperature (TEMPZ) were collected from each major drought event occurring during the period 1931–1985 for eight geographically dispersed climatic divisions within the contiguous United States. Average values of each index were calculated using data from month one, two, etc. to produce a time series of drought severity for each climatic division. Multiple-line graphs are used to present the time series.

Overlaying the graphs from any two climatic divisions reveals that there are no perfect matches in the temporal progression of drought severity within the sample. However, strong similarities in the patterns are evident among several climatic divisions, and the trends for certain pairs of drought indices (e.g. PHDI, PDSI) are largely consistent across all climatic divisions. Although there is some spatial variability in the strength of this relationship, the graphs show that during the initial months of an average drought event temperatures will be above normal and precipitation below normal. The longer these conditions persist, the greater the magnitude of meteorological and hydrological droughts.

RÉSUMÉ

Le but de base de cet article est d'examiner les schémas temporels simultanés d'événements naturels du type-sécheresse pendant les six premiers mois d'une sécheresse moyenne. La méthodologie employée dans cette étude a pour but d'aider la recherche de signaux sous-jacents ou bien de schémas typiques dans un échantillon de sécheresses. Les valeurs mensuelles du Palmer Moisture Anomaly Index (ZINX), du Palmer Drought Severity Index (PDSI), du Palmer Hydrologic Drought Index (PHDI) et les relevés standardisés de

précipitation (PREZ) et de température (TEMPZ) ont été réunis à partir de chaque sécheresse importante ayant eu lieu pendant la période allant de 1931 à 1985 dans huit divisions climatiques dispersées du point de vue géographique aux États-Unis. Les valeurs moyennes de chaque indice ont été calculées à partir des données du mois 1, 2, etc. afin de produire un schéma temporel de sévérité de sécheresse pour chaque division climatique. Des graphiques à plusieurs courbes sont utilisés pour présenter les schémas temporels.

La superposition des graphiques de deux divisions climatiques quelconques révèle qu'il n'y a pas de cohésion parfaite dans la progression temporelle de la gravité de la sécheresse dans l'échantillon. Cependant des similarités importantes dans les schémas sont évidentes parmi de nombreuses divisions climatiques et les tendances pour certaines paires d'indicateurs de sécheresse (e.g. PHDI, PDSI) sont largement consistantes à travers toutes les divisions climatiques. Bien qu'il y ait une variabilité spatiale en ce qui concerne la force de leur relation les graphiques montrent que durant les mois initiaux d'une sécheresse moyenne les températures seront en général au dessus de la moyenne et les précipitations en-dessous. Plus ces conditions persistent, plus la magnitude des sécheresses météorologiques et hydrologiques est grande.

1. INTRODUCTION

A general theme in the justification of drought research is that by increasing the base of knowledge of spatial and temporal characteristics of drought, more effective drought planning and management strategies can be developed (Yevjevich *et al.* 1978, Karl and Koscielny 1982, Wilhite 1983, Gregory 1986). Analyses which focus on the properties of drought nearly always employ some form of drought severity index. However, no universally acceptable definition or measure of drought exists at the current time. Instead, numerous definitions and measures of drought intensity geared to specific aspects of drought are used.

The drought types recognized by climatologists and others involved in drought research include, among others: agricultural, meteorological, hydrological, and forest-fire potential. The existing knowledge on temporal and spatial patterns of drought behavior is focused almost entirely on specific aspects of these individual drought types. Some studies which have dealt with multiple drought types (e.g. Doornkamp and Gregory 1980, Changnon *et al.* 1982) are unintegrated, treating each type as a separate entity. Analyses which have compared spatial and/or temporal patterns of drought as defined by two or more indices are either regionally focused (e.g. Oladipo 1985, Karl and Young 1987, Soulé 1988) or are primarily concerned with operational characteristics of the index (e.g. Karl 1986). Because of this lack of integration in drought studies, many aspects of the concurrent and lagged relationships among different drought types have yet to be examined.

The primary purpose of this study is to examine the temporal evolution of three measures of drought severity during the early stages of an average drought event for a sample of climatic divisions in the contiguous United States. Examining average drought conditions is a first step towards determining whether recurring temporal and/or spatial patterns of drought development exist in the United States. Guiding research questions for this study include: 1) what is the relationship and variation through time among three types of drought as measured by averaged drought severity indices, 2) what is the relationship and variation through time between averaged departures of precipitation and temperature from normal during the drought event, and 3) is there spatial variability in the temporal progression of drought severity, temperature, and precipitation as measured by the mix of indices?

2. DATA

All data used in this study were obtained from the 1986 version of the United States National Climatic Data Center's (NCDC) magnetic tape file TD9640 (NCDC 1986). The data used include monthly values of mean temperatures, total precipitation, the Palmer Moisture Anomaly Index (ZINX), the Palmer Drought Severity Index (PDSI), and the Palmer Hydrologic Drought Severity Index (PHDI) for eight climatic divisions in the United States for the period 1931–1985. The monthly values of all parameters are calculated by equally weighting all stations which record both temperature and precipitation within the boundaries of a climatic division (NCDC 1986). In addition, all data on this file have been corrected for the time-of-observation bias described by Karl *et al.* (1986).

The Palmer indices are water-balance based, matching demand for moisture (evapotranspiration) with supply (precipitation). The indices are designed so that values near zero represent normal moisture conditions, negative values represent moisture conditions below normal (drought), and positive values moisture conditions above normal (wet spells). Index values are generally in the range −7 to +7. Descriptions of drought (wet spell) severity associated with different ranges of the PDSI are shown in Table 1. Table 2 gives the classification for the ZINX and PHDI.

Because the Palmer indices respond to changes in supply and demand of moisture at different rates, they are often described as representing different types of drought. The ZINX has the fastest rate of response to changes in soil moisture of the three indices. It is described as a measure of agricultural drought, since rapid changes in soil moisture often have significant impacts on agricultural activities (Karl 1986). The PHDI responds slowly to changes in moisture supply and demand and is used as a measure of hydrological drought severity (Karl and Knight 1985). The PDSI has an intermediate rate of response to moisture changes and is generally referred to as a measure of meteorological or average drought severity (Palmer 1965). In addition to Palmer's (1965) original description, thorough explanations of the procedures used to calculate the indices are given by Alley (1984), Karl and Knight (1985) and Karl (1986).

In addition to the Palmer indices, standardized scores (z-scores) of monthly precipitation (PREZ) and average temperature (TEMPZ) were calculated. The PREZ provides for a measure of drought intensity based solely on

TABLE 1: Severity classes for the PDSI.

PDSI Value	Severity Class
≥ 4.00	Extremely wet
3.00 to 3.99	Very wet
2.00 to 2.99	Moderately wet
1.00 to 1.99	Slightly wet
0.50 to 0.99	Incipient wet spell
0.49 to −0.49	Near normal
−0.50 to −0.99	Incipient drought
−1.00 to −1.99	Mild drought
−2.00 to −2.99	Moderate drought
−3.00 to −3.99	Severe drought
≤ −4.00	Extreme drought

(after Palmer, 1965, p. 28)

TABLE 2: Severity classes for the PHDI and ZINX.

Approximate Cumulative Frequency (%)	PHDI Value	Severity Class	ZINX Value
≥ 96	≥ 4.00	Extreme wetness	≥ 3.50
90–95	3.00 to 3.99	Severe wetness	2.50 to 3.49
73–89	1.50 to 2.99	Mild to moderate wetness	1.00 to 2.49
28–72	−1.49 to 1.49	Near normal	−1.24 to 0.99
11–27	−1.50 to −2.99	Mild to moderate drought	−1.25 to −1.99
5–10	−3.00 to −3.99	Severe drought	−2.00 to −2.74
≤ 4	≤ −4.00	Extreme drought	≤ −2.75

(after NCDC, 1986, p. 3)

the deviation of precipitation from normal. Although there is no qualitative assessment of drought severity associated with PREZ values as there is with the PDSI, ZINX, and PHDI (Tables 1, 2), examining the monthly precipitation and temperature anomalies adds to the understanding of a total drought event. PREZ and TEMPZ were calculated using the STANDARD procedure of the Statistical Analysis System (SAS Institute 1985).

3. METHODS

The relationships examined in this study are based on the drought histories, as measured by the ZINX, PDSI, and PHDI, of eight climatic divisions in the contiguous United States (Figure 1). These eight climatic divisions represent the

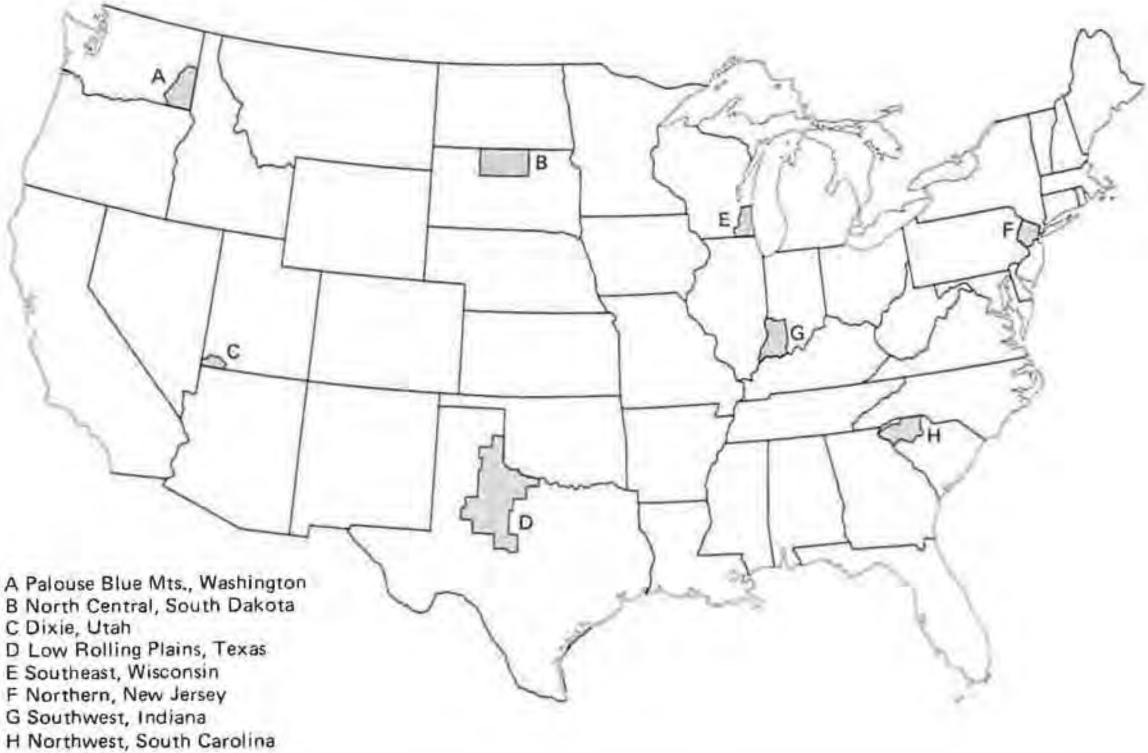

FIGURE 1. Locations of sample climatic divisions.

centroids of spatially homogeneous meteorological drought regions (Soulé and Meentemeyer 1989) and provide for a geographically dispersed sample across a variety of climate types.

For this study a major drought event was defined as any span of six or more consecutive months with any of the Palmer indices ≤ -1.0. Diaz (1983) gives a similar definition in which a string of six or more consecutive months with PDSI values ≤ -2.0 is considered a major drought. Relaxing the Diaz definition and including the fast responding ZINX and slowly responding PHDI in the formula results in a very liberal definition of a major drought event. By design, this definition ensures that the beginning and ending periods of long drought events are included in the analysis.

The following procedure was used for each of the eight sample climatic divisions. Monthly data for the PREZ, ZINX, PDSI, PHDI, and TEMPZ were extracted for each major drought. Average index values were then calculated using data from the first month of all major drought events. The procedure was repeated for the second through sixth months to produce a time series of average drought severity, precipitation, and temperature index values. The index values (and temperature/precipitation z-scores) were then plotted to produce a multiple-line graph displaying the progression of drought severity among the multiple drought types during the first six months of the average major drought event.

4. RESULTS

As shown in Table 3, the relationship between frequency and duration of major droughts is strongly negative ($r = -0.86$) across the sample. This relationship is influenced by the fact that as drought duration increases there is a reduced opportunity for large numbers of droughts to be recorded. The North Central, South Dakota climatic division experienced a small number of major droughts

TABLE 3: Drought frequency and duration for the eight sample climatic divisions.

Climatic Division	Number of Major Droughts (1931–85)	Average Length of Major Droughts (months)	Ratio: Number of Droughts/ Drought Length
NC S. Dakota*	8	29.5	0.27
SW Indiana*	11	19.8	0.55
LRP Texas	14	19.6	0.71
N. New Jersey*	13	16.2	0.80
NW S. Carolina	15	16.6	0.90
PBM Washington*	15	16.5	0.91
Dixie Utah	18	16.3	1.10
SE Wisconsin*	17	15.0	1.13

* Indicates that the climatic division was experiencing a major drought beginning prior to January, 1931. This first drought is not counted or involved in the calculation of drought lengths.

during the 55 year study period, but these droughts were of long duration. Since snow cover is not factored into the computational routines for the Palmer indices, the drought frequency and duration results may be slightly imprecise in high latitudes. However, the greater persistence of droughts in this portion of the interior United States is consistent with the findings of both Karl (1983) and Diaz (1983). At the opposite end of the sample are the Dixie, Utah and Southeast, Wisconsin climatic divisions, where major droughts occur more frequently, but have a much shorter average duration. Although they are located in different climatic regimes, the ratios of drought frequency to duration are nearly identical for these two climatic divisions.

It can be argued that all of the major droughts occurring in the eight sample climatic divisions during the study period were unique events, each related to a particular combination of synoptic meteorological controls. Considering that these droughts start and end in all seasons, and last from a minimum of six to upwards of one hundred months, this would seem to be a reasonable assumption. The methodology employed in this study was designed to help determine the underlying signals or patterns to these singular drought events during their early stages, as expressed through the mix of average drought severity, precipitation, and temperature index values.

Multiple-line graphs displaying the progression of average PREZ, ZINX, PDSI, PHDI, and TEMPZ values during the first six months of the (average) major drought event are presented for those climatic divisions in the western United States in Figure 2, and those divisions in the central and eastern states in Figure 3. Comparing the graphs from any two climatic divisions reveals that there are no perfect matches in the temporal progression of drought severity within the sample. However, strong similarities in the patterns are evident among several climatic divisions, and the trends for certain pairs of drought indices (e.g. PHDI, PDSI) are largely consistent across all climatic divisions.

The most consistent pattern across all graphs is the steady decline in PHDI and PDSI values through at least the fourth month of the drought event. The only deviation from this pattern comes in Northwest, South Carolina, where the PHDI increases in the second month before decreasing. The trend lines for these two indices are generally parallel during the early months of the (average) drought event, with PHDI values running slightly higher than PDSI values until the fourth, fifth, or sixth month of the drought. Another consistent pattern across all divisions is the low values of the PREZ and ZINX during the early months (one, two, and three) of the drought event. The trend lines for these two indices tend to be distinctly parallel throughout the six month period.

In Northwest, South Carolina, Low Rolling Plains, Texas, Palouse Blue Mountains, Washington, and Northern, New Jersey, there is an overall inverse relationship trend between TEMPZ and ZINX during the early portions of the average drought event (months 1–4), particularly the second to fourth months. The trends indicate that temperatures tend to be above normal and precipitation below normal during the initial months of the drought event (one and two),

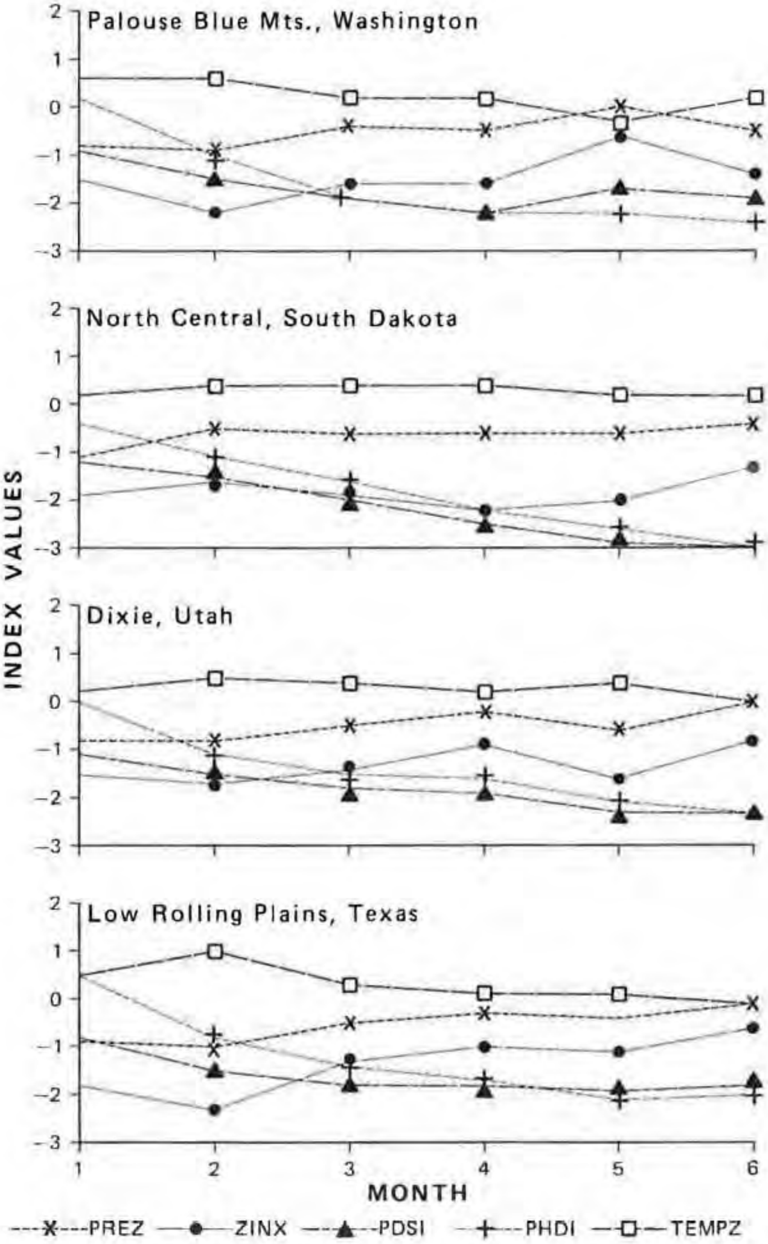

FIGURE 2. Temporal progression of average PREZ, ZINX, PDSI, PHDI, and TEMPZ values during the first six months of the average drought event for the sample climatic divisions in the western United States.

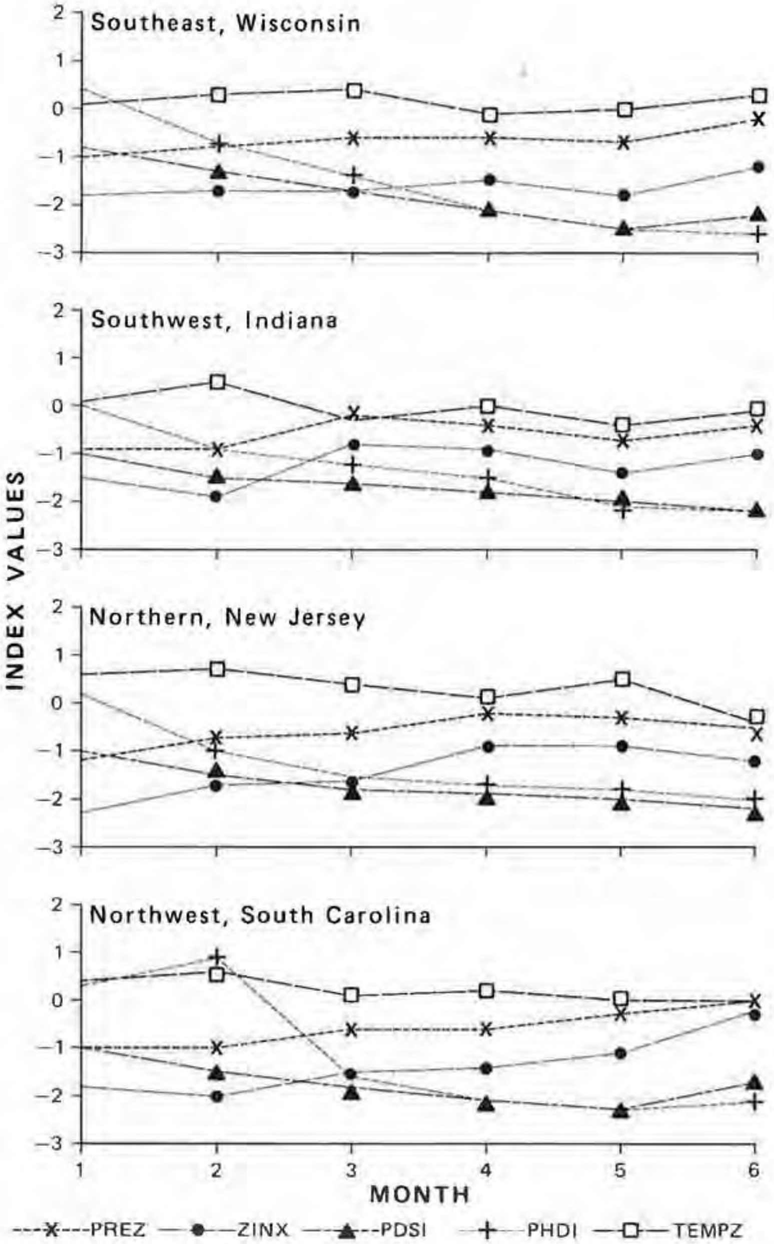

FIGURE 3. As in Figure 2, except for sample climatic divisions in the central and eastern United States.

followed by a gradual return to more normal conditions by the fourth month. In the fifth and sixth months the index values (PREZ, ZINX, TEMPZ) remain close to normal, although they do not follow a consistent trend across these four climatic divisions. For any given month, the magnitude of precipitation deviations from normal tends to exceed that of temperature at all locations.

In both Dixie, Utah and North Central, South Dakota there is little temporal variability in average temperature and precipitation index values. Temperatures remain slightly above and precipitation slightly below normal throughout the first six months of the drought event. In Southeast, Wisconsin and Southwest, Indiana there is an undulating pattern to TEMPZ and PREZ across the six months. Temperature and precipitation patterns in Indiana are clearly the most anomalous of the eight locations examined in this study, especially in months three to six. At all other locations TEMPZ is above normal through at least the first three months, and PREZ does not tend to approach normal until at least the fourth month.

The closest match in overall patterns is between Northwest, South Carolina and Low Rolling Plains, Texas, the two climatic divisions with the most southerly location. Another pair with closely matched overall patterns are the North Central, South Dakota and Southeast, Wisconsin climatic divisions. Similarity of the drought patterns at these climatic divisions is interesting considering that the average length of major droughts in North Central, South Dakota is almost twice that of Southeast, Wisconsin (Table 3). The magnitude of meteorological and hydrological drought severity in North Central, South Dakota during the fifth and sixth months is the largest among the sampled divisions. The consistency of average precipitation deviations during the drought event is reflective of the fact that droughts are more persistent in the northern Great Plains than in any other part of the country. Furthermore, the consistency of precipitation deviations through time in this average drought event would tend to support the notion of a positive feedback system for drought in this region.

5. SUMMARY

Despite the variability in the overall drought history across the sample (i.e. frequency, duration), the majority of inter-site differences in the average drought patterns are small, with the indices behaving in a similar fashion through time across the United States. For all locations, the greatest similarities in temporal development patterns are between the PHDI and PDSI, the two indices with the slowest rate of response to changes in supply and demand for moisture. While the temporal development patterns of the faster responding ZINX and PREZ are more erratic both within and between sites, the general trend for these indices shows a gradual return towards more normal moisture conditions by the sixth month of the drought. At all locations, temperature tends to be above normal and precipitation below normal during the initial months (one to three) of the average drought event. Although there is considerable spatial variability in this relationship, it is clear that the longer these conditions persist, the greater the magnitude of meteorological and hydrological droughts.

The development of General Circulation Models in meteorology has not yet reached the point where it is possible to produce accurate long-term forecasts. The ability to prepare for future droughts is therefore largely determined by an understanding of drought characteristics based on the climatic record. The methodology developed for this study involves the synthesis of drought information from selected series of drought events affecting a given region. Application of this methodology makes it possible to look for the underlying temporal patterns of drought development within the region. By modifying and expanding this methodology it should be possible to identify recurring temporal and/or spatial patterns of drought development within the United States.

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Geographical Patterns of Nocturnal Precipitation of the Canadian Prairies During the Growing Season

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ABSTRACT

A preliminary investigation of hourly precipitation data from selected stations on the Canadian Prairies was conducted in order to examine whether there was any diurnal pattern. Data for the growing seasons of 1965–74 were analysed. Precipitation amounts and number of hours reporting precipitation were greater at night than during the day at a majority of the stations.

RÉSUMÉ

Afin d'examiner la répartition temporelle des précipitations dans les Prairies canadiennes, nous avons étudié les données horaires de précipitation. Nous avons conclus que, durant la période 1965–1974, les quantités de précipitation et ses fréquences horaires, lors de la saison de croissance, étaient plus élevées la nuit que le jour à plusieurs stations météorologiques des Prairies.

1. INTRODUCTION

Several studies deal with the annual, seasonal or even daily characteristics of precipitation in the Canadian Prairies (Hopkins and Robillard, 1964; Longley, 1974; Chakravarti, 1976; Dey and Chakravarti, 1976; Dey, 1982). But the author is unaware of a detailed analysis of the *diurnal regime of precipitation* in the Canadian Prairies. Such a study could prove useful, for growing-season rainfall during the night is more available to crops because of reduced evaporation from soil and plant surfaces, allowing greater infiltration into the soil.

Kincer (1916) examined day-time and night-time precipitation and their economic significance. Kincer (1922) observed that two areas in the continental United States – the central Great Plains and southern Arizona – received 60 to 65 percent of their total summer rainfall during the night. Bleeker and Andre (1951), Sangster (1958), and Wallace (1975) have discussed the reasons for the nocturnal maximum of rainfall over parts of the United States. Trewartha (1981:293) noted that, although maximum shower activity over the Great Plains is normally expected during the hours of greatest surface heating, the region has a nocturnal maximum during the warmer season. The objective of this paper is to examine whether the Canadian Prairies, as part of the Great Plains, receive more of their growing-season precipitation during the night than during the day.

2. STUDY AREA AND DATA

The study area (Figure 1) comprises the agricultural southern half of the prairie provinces of Canada where diurnal variations in precipitation may have significant impacts on crop yields. Across this area there is a relatively uniform network of weather stations recording hourly precipitation amounts and events. From these, twenty stations (Figure 1) with a continuous record of hourly precipitation were selected to provide even coverage of the study area. Four of the stations (Jasper, Rocky Mountain House, Pincher Creek and Indian Bay) lie just outside the area, and were not used in the statistical testing described in sections 3.1 and 3.2, but were deemed useful in defining boundary conditions. Hourly precipitation data for the period 1965 to 1974, already available to the author from previous research (Chakravarti, 1976), were used for this preliminary study.

3. PRECIPITATION PATTERNS

Earlier researchers (Kincer, 1922; Trewartha, 1981; Balling, 1982) have defined locations with more than 50 percent of precipitation occurring between 2000 LST and 0800 LST as having a nocturnal maximum of precipitation. For convenience and for comparative purposes, the same definition of nocturnal precipitation maximum has been used here. Two main features are examined, viz. (a) more than 50 percent of the total amount, and (b) more than 50 percent of the total number of hourly reports of precipitation occurring between 2000 and 0800 Local Standard Time.

3.1 *Precipitation Amounts*

Table 1 shows that a number of the selected stations recorded more than 50 percent of their growing-season precipitation for the period 1965–74 falling between 2000 and 0800. In terms of the monthly groupings, the maximum percentage was 78.36 for Brooks in southern Alberta for August, and the minimum 34.95 at Scott, Saskatchewan for September. The averages for the group of stations in Alberta and Manitoba show that more than 50 percent of the total precipitation occurred at night in each monthly grouping May to August, and also over the entire growing season (Table 1). Some individual stations in each province, however, had one or more monthly groupings with less than 50 percent of the precipitation occurring at night.

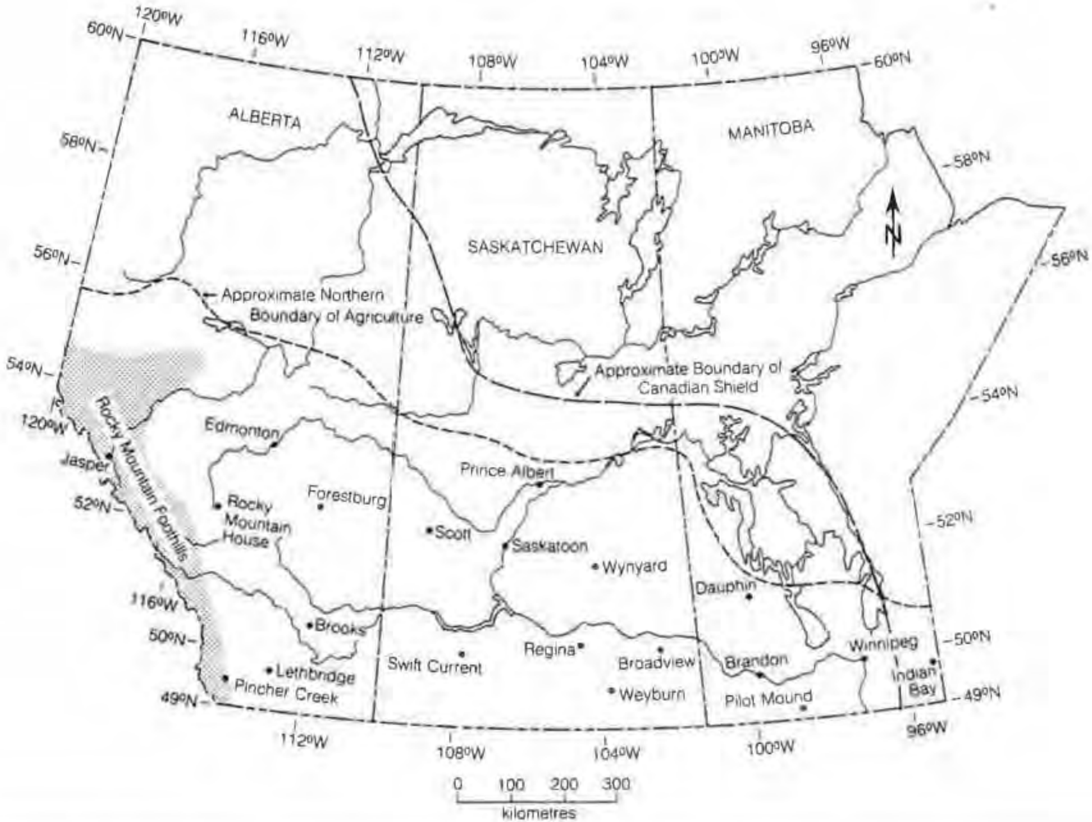

FIGURE 1. The study area and the selected stations.

TABLE 1. Percentage of total precipitation falling between the hours of 2000 LST and 0800 LST at selected stations in the Canadian Prairies, 1965–1974.

	May	June	July	Aug.	Sept.	Growing Season
ALBERTA						
Brooks	51.06	51.36	65.62	78.36	58.40	59.14
Edmonton	53.42	49.82	46.67	50.22	44.36	50.62
Forestburg	44.66	42.85	62.38	55.07	42.85	52.46
Jasper	52.99	50.77	49.97	48.21	54.26	50.09
Lethbridge	57.26	61.76	55.66	48.26	56.89	57.51
Pincher Creek	55.36	51.01	52.63	46.40	47.35	50.60
Rocky Mtn. House	47.59	55.90	39.39	55.07	40.89	47.86
Average	51.76	51.92	52.76	54.51	49.29	52.61
SASKATCHEWAN						
Broadview	43.61	51.99	43.26	50.99	48.55	48.02
Prince Albert	54.93	44.48	51.56	49.97	51.62	49.62
Regina	49.63	42.16	55.48	51.67	51.66	49.60
Saskatoon	45.59	47.48	62.05	40.45	47.21	49.72
Scott	44.28	44.34	55.56	50.29	34.95	47.63
Swift Current	46.17	56.96	49.66	43.35	51.96	50.41
Weyburn	52.16	47.94	64.09	51.12	46.75	52.59
Wynyard	49.60	43.62	48.26	61.01	63.72	51.84
Average	48.25	47.37	53.74	49.86	49.55	49.93
MANITOBA						
Brandon	51.17	52.15	54.09	67.10	50.88	54.94
Dauphin	49.51	65.00	50.30	41.94	54.63	53.23
Indian Bay	48.22	41.64	46.80	58.81	44.22	48.00
Pilot Mound	55.31	50.77	58.90	54.89	58.96	55.95
Winnipeg	59.46	52.07	57.13	54.06	55.51	55.87
Average	52.73	52.33	53.44	55.36	52.84	53.60

TABLE 2. Paired difference t-test for the amount of precipitation occurring between the hours of 2000 LST and 0800 LST for the group of stations.†

	May	June	July	Aug.	Sept.	Growing Season
Alberta Stations	1.10	0.74	1.25	1.90**	0.48	2.44**
Saskatchewan Stations	−1.23	−1.58	1.57*	0.16	0.06	−0.01
Manitoba Stations	1.70*	1.43	2.58**	1.05	2.85**	4.07**
Prairies – Total Group of Stations	1.08	0.43	3.08**	1.73*	1.40*	3.29**

† Excludes non-Prairie stations: Jasper, Rocky Mtn. House and Pincher Creek, Alta., and Indian Bay, Man. (see Table 1).

* $p < 0.10$

** $p < 0.05$; statistically significant

A paired difference t-test was used to test the hypothesis that the day-time (0800–2000) and the night-time (2000–0800) amounts were equal. The t-test was used for the group of stations in each province on the cumulative differences between night-time and day-time precipitation amounts for each monthly grouping separately and also for the growing season. A p-value of 0.05 or less was used as the criterion for statistical significance.

The results are shown in Table 2. Over the entire growing season, the hypothesis was rejected at the .05 level for the groups of Alberta and (especially) Manitoba stations. Thus night-time precipitation was significantly greater than day-time precipitation, and this was also the case with the entire group of selected prairie stations. Some of the monthly groupings also showed a statistically significant nocturnal maximum, and others approached significance (see Table 2).

The spatial pattern of the proportion of nocturnal precipitation during the growing season over the period 1965–74 is shown in Figure 2. Large areas of the prairies experienced a nocturnal maximum, but in west-central Alberta and north-central Saskatchewan night-time precipitation was less than 50 percent of the total.

3.2 *Hourly Observations of Precipitation*

As with the amount of precipitation, more than 50 percent of total hourly observations with precipitation was also recorded between 2000 and 0800 by most of the selected stations in Alberta and Manitoba (Table 3). Although there were only three stations in Saskatchewan recording more than 50 percent of the hourly observations at night, the other three were very close to 50 percent. The groups of all selected stations in a) Alberta, and (b) Manitoba show most monthly groupings and the entire growing season with more than 50 percent of the total hourly observations with precipitation coming between 2000 and 0800 hours, similar to the precipitation amounts.

As with the amounts, a paired difference t-test was used to assess the hypothesis that day-time and night-time hourly observations with precipitation were of equal frequency. The results of this test were broadly similar to those for precipitation amounts, with the station groupings for Alberta, Manitoba and the entire region for the growing season as a whole achieving statistical significance and demonstrating a nocturnal maximum of frequency of hours with precipitation (Table 4). Some of the monthly groupings also achieved statistical significance.

The spatial pattern of the nocturnal percentage of hours reporting precipitation was also plotted (Figure 3). There is a general similarity between the areas with nocturnal precipitation maxima (Figure 2) and those with nocturnal maxima of precipitation frequency (Figure 3). Again, there are areas of low values in western Alberta and central Saskatchewan; and in southern Alberta, southwestern Saskatchewan and southern Manitoba, there are nocturnal maxima, with frequency from 52 to 56 percent (Figure 3).

An attempt was also made to determine if there were any concentrations of precipitation during particular hours of the day during the

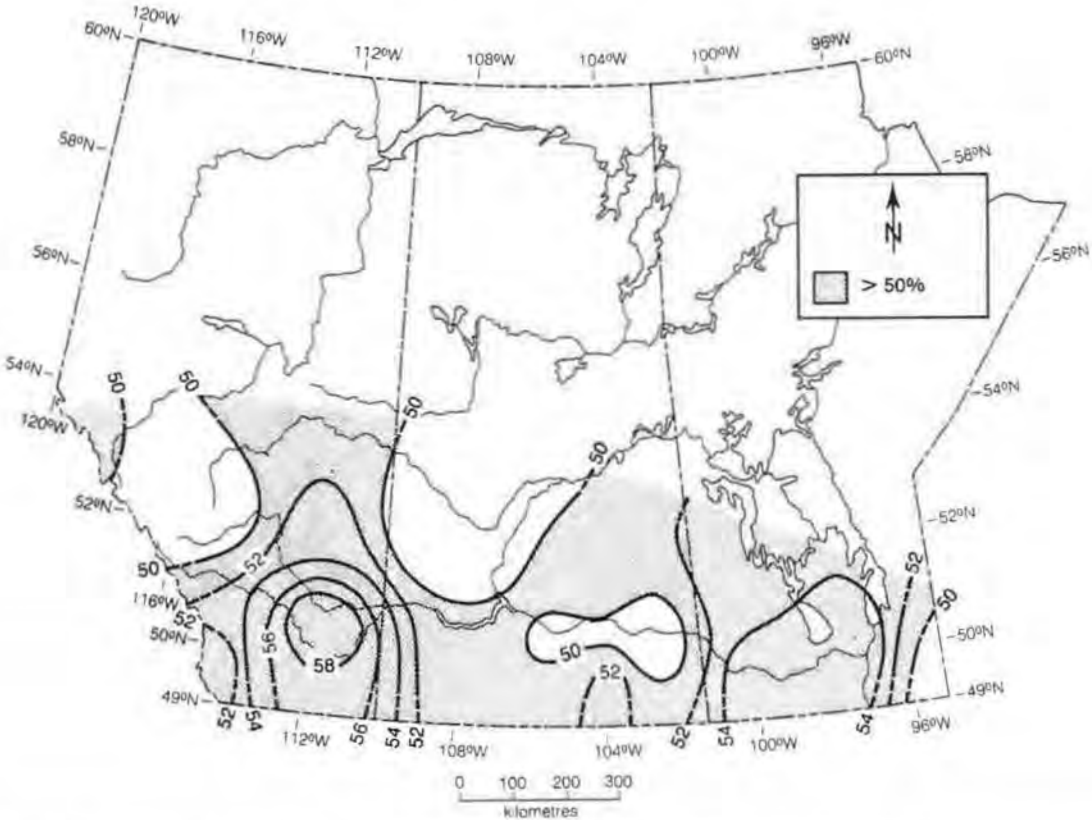

FIGURE 2. Percentage of total precipitation received between the hours of 2000 LST and 0800 LST during the growing season.

TABLE 3. Percentage of number of hourly observations reporting precipitation which fall between the hours of 2000 LST and 0800 LST at selected stations in the Canadian Prairies, 1965–1974.

	May	June	July	Aug.	Sept.	Growing Season
ALBERTA						
Brooks	50.00	54.09	64.01	67.85	54.46	57.18
Edmonton	51.73	47.63	47.16	55.32	43.89	49.05
Forestburg	43.79	44.85	55.62	58.22	50.00	51.29
Jasper	53.27	55.01	50.00	59.23	54.46	53.99
Lethbridge	54.61	57.14	64.35	50.21	54.88	56.16
Pincher Creek	54.33	51.22	51.74	50.00	48.62	51.24
Rocky Mtn. House	46.06	52.29	45.58	49.73	46.30	48.46
Average	50.54	51.75	54.07	55.79	50.37	52.48
SASKATCHEWAN						
Broadview	47.69	52.87	50.70	52.06	50.96	50.91
Prince Albert	54.85	47.22	47.40	41.62	48.71	47.31
Regina	47.22	49.05	47.03	52.66	50.32	49.09
Saskatoon	55.31	51.79	51.80	41.58	46.90	49.79
Scott	54.85	47.02	47.64	41.62	48.71	48.66
Swift Current	47.38	52.00	42.85	53.82	47.32	49.06
Weyburn	53.20	50.15	48.12	48.03	52.34	50.69
Wynyard	51.12	52.34	46.56	53.17	53.31	51.29
Average	51.45	50.31	47.76	48.07	49.82	49.60
MANITOBA						
Brandon	53.78	50.38	51.44	58.54	52.40	53.18
Dauphin	52.13	52.86	47.46	48.10	51.16	50.39
Indian Bay	45.39	43.76	43.76	48.49	51.06	46.50
Pilot Mound	55.53	52.55	55.44	56.82	52.44	54.61
Winnipeg	53.08	45.89	53.29	54.73	50.98	51.55
Average	51.98	49.09	50.28	53.34	51.61	51.25

TABLE 4. Paired difference t-test for the frequency of precipitation occurring between the hours of 2000 LST and 0800 LST for the group of stations.†

	May	June	July	Aug.	Sept.	Growing Season
Alberta Stations	0.59	0.56	1.53	2.72**	0.11	2.33**
Saskatchewan Stations	0.80	0.25	−2.46	−1.01	0.48	−0.89
Manitoba Stations	4.13**	0.22	1.01	1.97*	5.30**	3.60**
Prairies – Total Group of Stations	2.30**	0.70	0.53	1.02	1.10	2.34**

† Excludes non-Prairie stations: Jasper, Rocky Mtn. House and Pincher Creek, Alta., and Indian Bay, Man. (see Table 1).

* $p < 0.10$

** $p < 0.05$; statistically significant

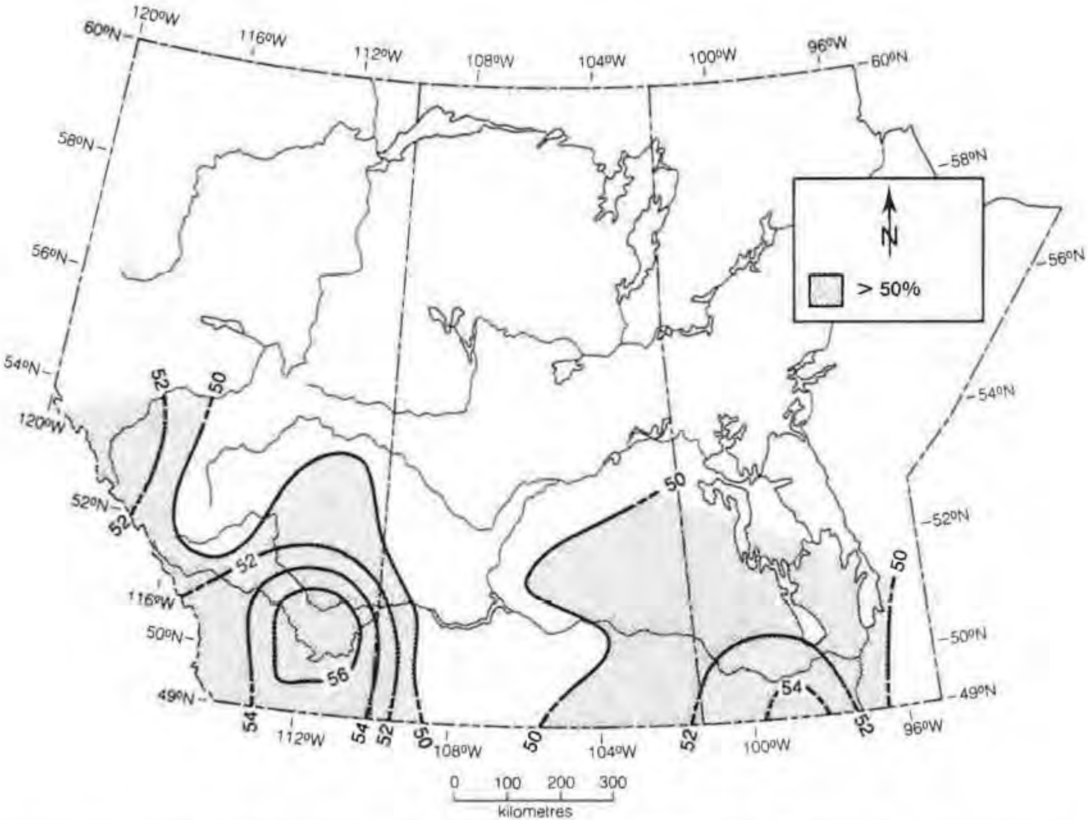

FIGURE 3. Percent frequency of hourly observations with precipitation falling between the hours of 2000 LST and 0800 LST during the growing season.

growing season. A few stations in Alberta and Manitoba showed a higher frequency of precipitation around midnight and/or early in the morning. But the pattern appeared to be random.

4. CONCLUSIONS

This preliminary analysis of 1965–74 data indicates that large areas in the important agricultural region of the Canadian prairies have greater amounts and frequency of precipitation at night than in the day during the growing season. This is a significant characteristic which contributes towards more effective precipitation for the environment in general and farming in particular.

Furthermore, occurrence of nocturnal precipitation in the Canadian prairies appears to be a northern extension of a similar pattern observed in the Great Plains in the United States. These nocturnal maxima of both the amount and the frequency of precipitation cannot result from the random convectional showers common during afternoons on the heated interior plains. As observed in the United States, such nocturnal patterns may be related to organized or recurrent surface or upper atmospheric flow patterns (Bleeker and Andre, 1951; Sangster, 1958). A further detailed analysis of precipitation patterns over a longer period of time and an explanation of this nocturnal maximum should be very useful not only in forecasting precipitation but also in contributing to better understanding of diurnal rainfall regimes during the growing season.

ACKNOWLEDGEMENTS

The author is grateful to Bonnie Magill and Elaine Wheaton for help in data collection and tabulation, Anne Patterson for computer work and mapping, Keith Bigelow for drafting the figures and to Dorothy Young for word processing.

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News and Comments
Nouvelles et commentaires

RECENT PUBLICATIONS

Florida Weather by Morton D. Winsberg.
University of Central Florida Press, Orlando, 1990: 171 pp. ISBN 0-8130-0989-8. This readable treatment of Florida's climate is available in paperback at U.S. $9.95 from: University Presses of Florida, 15 NW 15th Street, Gainesville, FL 32603, U.S.A.

Atlas of the Surface Heat Balance of the Continents by Dieter Henning.
Gebrüder Borntraeger, Stuttgart, 1989: 402 pp., in English. ISBN 3-443-01025-3. Available at DM 178, – in 19 × 29 cm page size from E. Schweizerbart'sche Verlagsbuchhandlung (Nägele u. Obermiller), Johannesstrasse 3A, D-7000 Stuttgart 1, W. Germany.

Climatological Study of Temperature Inversion Layers in the Northern Hemispheric Troposhere by F.H. Liu.
Climatological Notes 39, Institute of Geoscience, University of Tsukuba, Japan, 1990: 80 pp., in English. ISSN 0388-0206.

Climat et Santé: Cahiers de bioclimatologie et biométéorologie humaines.
Cahiers apériodiques lancés en 1989, dont on a reçu les deux premiers numéros dans le cadre d'une échange avec le Bulletin climatologique, disponible du: Groupement de Recherche "Climat et Santé", G.D.R. 102, Faculté de Médecine, 7 Boulevard Jeanne d'Arc, F-21033 DIJON Cedex, France. Tout en français, le no. 1 comprend huit articles et notes, soit 145 pages, et le no. 2 en a huit aussi, pour 161 pp.

CORRIGENDUM

Australian Workshop on Bushfire Meteorology and Dynamics

Thanks to Marty Alexander for pointing out that on the third line of p. 136 of Vol. 23(3), December 1989, the word "above" should be substituted for the word "below". Information on this Workshop is available from Dr. Tom Beer, CSIRO Division of Atmospheric Research, Private Mail Bag No. 1, Mordialloc, Victoria 3195, Australia, and not from the Northern Forestry Centre in Edmonton.

List of Referees / Liste des arbitres

The previous list was published in Volume 22(3), December 1988. This new list covers most of 1988, 1989 and part of 1990. *Climatological Bulletin* sincerely appreciates the time and care taken by the referees listed herein.

La dernière de ces listes a été publiée en Volume 22(3), décembre 1988. Cette nouvelle liste est consacrée à la plupart de l'an 1988, 1989 et une partie de 1990. Le *Bulletin climatologique* apprécie sincèrement le temps et les efforts des arbitres soulignés ici.

D.A. Bernachi
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A.J. Brazel

A.K. Chakravarti
S.J. Cohen

J.A. Dyer
P. Dzikowski

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R.J. Fletcher
D.B. Frost

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