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Measurement and analysis of small-scale convective storm rainfall variability

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Abstract

For large-scale catchment hydrology, the crucial importance of a good estimate of spatial rainfall variability is generally admitted. However, the assumption of uniform rainfall is still applied for small areas, whether they are studied as individual catchments or represent an elementary area in a distributed model. To investigate the validity of this assumption, an experiment was conducted in a small catchment (4.4 ha) in the semiarid USDA-ARS Walnut Gulch Experimental Watershed. Measurements were made with recording and non-recording raingages as well as vectopluviometers for a range of events during the 1990 monsoon season (July–September). Geostatistical analysis of the data indicated the presence of first-order drift with corresponding rainfall gradients ranging from 0.28 to 2.48 mm per 100 m with an average of 1.2 mm per 100 m. These gradients represent a 4–14% variation of the mean rainfall depth over a 100 m distance. Given these observations, the assumption of spatial rainfall uniformity in this and similar convective environments at the small watershed scale of 5 ha appears to be invalid.

1. Introduction

The spatial variability of rainfall plays an important role in the process of surface runoff generation, yet the assumption of uniform rainfall is still applied in modeling the hydrological behavior of small watersheds. In a region characterized by convective thunderstorms Goodrich (1990) noted that two raingages approximately

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300 m apart often provided significantly different estimates of rainfall depth and intensity. When these rainfall measurements were used as inputs to a distributed rainfall–runoff model on three small catchments (0.4–4.4 ha), significant sensitivity of the model to spatial variability of rainfall input was noted. The question then arises: is the assumption of rainfall field homogeneity at such a small scale valid and what role does precipitation measurement uncertainty play in interpreting spatial rainfall variability? This is of prime importance in assessing the reliability of small watershed runoff models.

The overall objective of this study is to better understand how spatial rainfall variability impacts the performance of runoff models for small catchments (< 5 ha). Four objectives must be addressed to provide this understanding.

(1) Assess precipitation measurement uncertainty resulting from gage type, calibration, data reduction and placement.

(2) Assess the impacts of wind on precipitation observations.

(3) Evaluate, at a very small catchment (< 5 ha) and temporal (< 5 min) scale, if sufficient spatial and temporal rainfall variability exists to have a significant effect on the estimate of areal precipitation over the catchment.

(4) Assess the impacts of the above three factors on runoff prediction accuracy?

In this paper an experimental approach was utilized to address the first three objectives. The fourth is explored in a companion paper (Faurès et al., 1995, this issue).

The Walnut Gulch Experimental Watershed, near Tombstone, Arizona provides an ideal location for such an experiment (Fig. 1(a)). At a larger scale (from 1 to

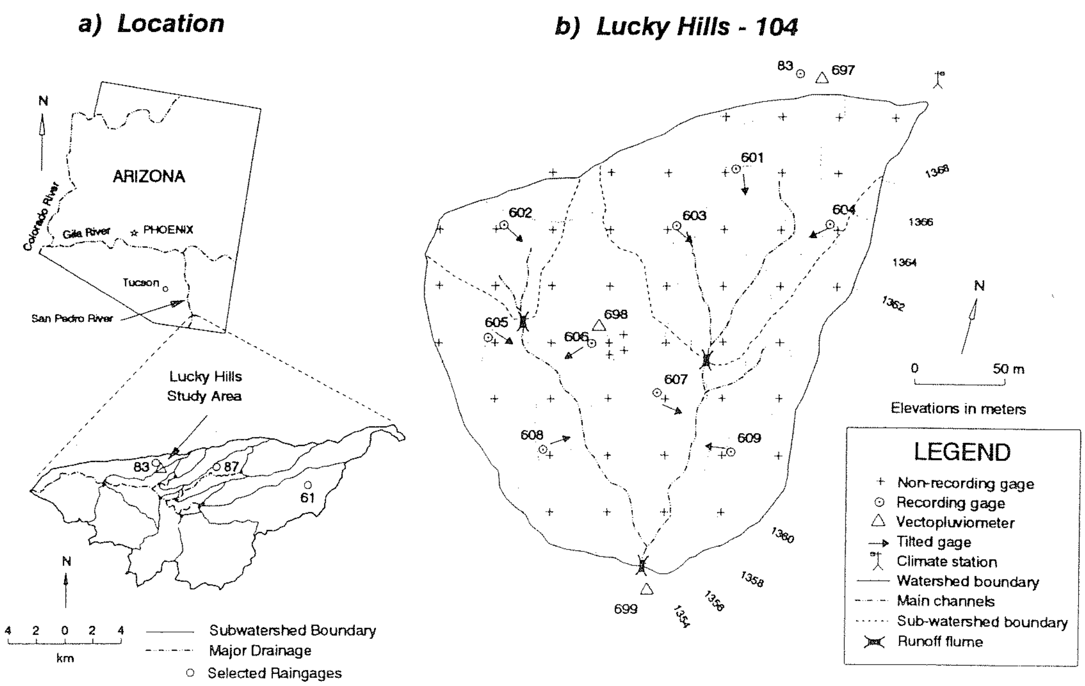

Fig. 1. (a) USDA-ARS Walnut Gulch Experimental Watershed location map, and (b) Lucky Hills-104 small-scale experimental network.

150 km^2), its hydrology is well known and numerous studies provide information concerning the spatial and temporal aspects of convective storms (e.g. Renard et al., 1993). At the smaller scale, a body of hydrologic and erosion analyses and modeling has been conducted in the Lucky Hills watersheds (0.4–4.4 ha, Fig. 1(b)) using the Water Erosion Prediction Project (WEPP) model (Parker, 1991; Tiscareno, 1991; Van der Zweep, 1992) and KINEROS (Goodrich, 1990; Woolhiser et al., 1990).

2. Background

Although the spatial variability of rainfall and associated measurement accuracy have been studied for some time (Osborn and Reynolds, 1963; Osborn and Keppel, 1966; Drissel and Osborn, 1968; Osborn et al., 1972, 1979; Rodda, 1967, 1973), they take on special importance when coupled with complex distributed rainfall–runoff models. This is particularly true at a very small scale (< 100 ha), where minimal variations of rainfall induce important changes in model response owing to infiltration model sensitivity and short catchment response times. Given the relatively short response times involved in urban catchments the importance of spatial rainfall variability for storm drainage design has been well documented (Schilling, 1984, 1991; Niemczynowicz, 1989, 1991).

Three sources of error have been identified by Dreaver and Hutchinson (1974) as well as Freimund (1992) and were investigated in this study: (1) systematic spatial and attitudinal variations; (2) systematic measurement error; (3) random measurement and sampling errors (owing to representation of an areal value by some function of a point value).

2.1. Systematic spatial and attitudinal variability of storms

Spatial variability of precipitation has been measured by means of radar, raingage networks or combinations of the two. In regions where thunderstorms prevail, dense raingage networks are necessary as convective storms are known to exhibit significant spatial and temporal variability (Osborn, 1967; Sharon, 1972; Osborn and Lane, 1981; May and Julien, 1990). Numerous descriptions of spatial rainfall variability over catchments of 1 km^2 to several hundred square kilometers are contained in special symposia devoted to the problems of urban storm drainage (Balmer et al., 1984; Iwasa and Sueishi, 1990; Niemczynowicz and Sevruk, 1991) and precipitation measurements (Sevruk, 1989a; Sevruk and Lapin, 1993). A good review of the needs for spatial and temporal rainfall analyses is provided by Berndtsson and Niemczynowicz (1988). Using one of the densest gage networks found in the literature (14 raingages in a 36 ha catchment) Ambroise and Aduizian-Gerard (1989) noted significant attitudinal variability of rainfall in a mountain catchment (20% mean slope) which was primarily attributed to topographically controlled wind direction.

2.2. Systematic measurement errors

Systematic errors can be of various types: water loss during the measurement, adhesion loss on the surface of the gage, and raindrop splash from the collector (for a good review, see Sevruk, 1989b, 1993). These errors are cumulative and lead to an underestimation of the actual amount of rain. Raindrop splash into the gage decreases with height of the gage orifice above the ground and will lead to an over-estimation. However, conflicting estimates of the effect of insplashing, the height at which it is important and the effect of various surrounding ground cover are present in the literature (Gold, 1931; Ashmore, 1934; Kurtyka, 1953; Green, 1970; Martinez, 1979; Cosandey, 1989).

A more serious systematic error is the measurement deficiency due to wind as the presence of the gage induces a disturbance of the horizontal component of the airflow (Larson and Peck, 1974; Green, 1970; Sevruk, 1989c; Zhi-Bing and Yun-Tiao, 1989). This systematic error also has a random component as the degree of underestimation depends on the random variations of wind speed, direction, and drop size (Freimund, 1992; Neff, 1977).

Another source of error can be found in the data reduction process. For weighing-type recording raingages, cumulative rainfall is generally recorded on paper charts, which must then be digitized. For this type of gage Chery and Kagan (1975) and Freimund (1992) note that the idealized resolution for a 1:1 depth scale to rain depth scale is 0.25 ± 0.25 mm (0.01 ± 0.01 in) but the practical processing resolution of the depth scale is 0.5 ± 1.0 mm (0.02 ± 0.04 in).

2.3. Random measurement and sampling errors

Random errors, resulting from imprecision in measurement, are usually small and compensating (Linsley et al., 1982). However, a more important error, related to the nature of the sample location, is the aspect of rainfall catch variability as influenced by local topography (Burns, 1953; Ambroise and Aduizian-Gerard, 1989). For a fixed wind speed and direction the rainfall measurement errors caused by topographic effects could logically be considered as systematic errors but with the imposition of a random wind field they were assumed random in this study.

Standard gages, adopted by the national service of a country responsible for rain-fall measurement are typically vertical gages (horizontal rim), installed at a fixed height above the ground. The use of this type of gage has led many to a misunder-standing of rainfall measurement when dealing with sloping ground. Peck (1973) concluded that "the basic question is to determine what we really wish to measure. If it is to compare the precipitation in different regions, then the vertical gage must be used. If an estimate of hydrologic precipitation is needed, then tilted gages may be used." What Peck (1973) calls hydrologic precipitation is the amount of water that actually reaches the ground. On a slope, this amount is not the same as the amount of rain falling on a horizontal surface unless wind speed is nil. This was proven theoretically by Fourcade (1942) and Hamilton (1954).

Hamilton (1954) and Sharon (1980) derived equations relating hydrologic precipitation to conventional gage data. This computation can be performed only if the drop inclination and azimuth are known. These quantities are often difficult to obtain. In a study in the Negev desert (Israel), Sharon (1980) reports rainfall inclinations between 40° and 60° in storms with wind speed of 10 m s^{-1}. The authors concluded that in order to measure the hydrologic precipitation, gages oriented with their rim parallel to the ground are necessary if no accurate way of measuring drop trajectories is provided. These gages should provide an unbiased estimate of the watershed hydrologic precipitation, provided the wind field is not highly variable and the measurements are used to represent the precipitation on topographic facets of uniform slope and exposure (Storey and Wilm, 1944).

From the foregoing review, it is apparent that little is known of the temporal and spatial rainfall variability and its impact on runoff prediction at scales less than 5 ha. In addition, previous experiments have shown that measurement errors may be a significant fraction of the total rainfall amount. To investigate small-scale rainfall variability and the impacts of measurement error the following approach was undertaken.

3. Approach

3.1. Network design

Measuring the small-scale variability of rainfall can be approached experimentally in two ways. A first possibility is to install a very dense network of vertical gages uniformly distributed over the study area with no consideration for catchment topography. This option should give good results if the variations resulting from convective storm physical properties are important and provided that the joint effect of wind and the hillslope are of little importance. Any method of spatial interpolation could be used to compute areal rainfall in this case.

The second solution would be, as suggested by Shih (1982), to divide the watershed in hydrological homogeneous areas with uniform vegetation, elevation, soil, physiography, slope and orientation. In this case, one or two gages should be installed in each zone with their rims parallel to the mean slope. Areal rainfall computed by interpolation is not justified in this case. Instead, the measures obtained from one raingage should be applied to the entire hillslope zone it represents (dimensions on the order of 30–50 m). This second solution should provide more representative results in a situation where wind-driven rainfall and catchment slope induce important variations in hydrologic precipitation.

Taking this information into account, and in an attempt to combine the two approaches, the raingage network described in Table 1 and Fig. 1(b) was designed and installed in Lucky Hills catchment no. 104 (LH-104). The four collocated vertical non-recording raingages (CNRG) were placed in the center of the catchment (site 606, Fig. 1(b)) to estimate rainfall measurement and microscale exposure variability for this gage type. One Time Domain Reflectometry (TDR) probe was installed at each recording gage site to measure the average soil moisture over the top 15 cm of soil to

Table 1
Raingage network summary

Raingage type	Brand	Number	Location (Placement)	Orifice dia. (cm) Area (cm^2)	Orifice elev. above ground (cm)	Depth resolution (mm)
Non-recording (cylindrical)	Taylor 5 inch gage	48	30×30 m grid (vertical)	4.52 16.05	30	0.07 ± 0.15[a] 0.15 ± 0.30[b]
Weighing recording (cylindrical)	Belfort (5-780) dual traverse	9	At center of nine areas of homogeneous slope and orientation (vertical)	20.32 324.29	100	0.25 ± 0.25[c] 0.5 ± 1.0[d]
Non-recording (cylindrical)	Taylor 5 inch gage	9	Collocated with recording gages (tilted)	4.52 16.05	30[e]	0.07 ± 0.15[a] 0.15 ± 0.30[b]
Non-recording (cylindrical)	Taylor 5 inch gage	4	Collocated with recording gage 606 in a 2×2 m area (vertical)	4.52 16.05	30	0.17 ± 0.15[a] 0.15 ± 0.30[b]
Vectopluviometer (cylindrical)[f]	Locally fabricated	3	Crest, center and watershed outlet	6.67 34.94	100	0.07 ± 0.15[a] 0.15 ± 0.30[b]

[a] Measurements made by poring raincatch into a 25 ml graduated cylinder for total depths from 0.0 to 15.0 mm.
[b] Measurements made by poring raincatch into a 100 ml graduated cylinder for total depths from 15.1 to 60.0 mm.
[c] Chart resolution.
[d] Practical processing (postdigitizing) resolution (Chery and Kagan, 1975).
[e] As measured perpendicular mean ground slope.
[f] One vertical measurement and four measurements in each of the principal directions at a 45° angle from vertical.

determine prestorm soil moisture conditions for runoff modeling (Dalton and Van Genuchten, 1986). Wind data were obtained from an automatic climatic station situated at the northern edge of the watershed.

The small non-recording gages (NRG) used in this study (Table 1) were installed with their rim 30 cm above the ground as Sharon (1980) demonstrated the superiority of a lower, small orifice (2.92 cm diameter) gage in comparing pit gages (16 cm diameter) with anti-splash guard, to a standard Hellman-type recording raingage (16 cm diameter) at 1 m in height used in that study. The lower NRG underestimated pit gage data by 9.3% while the 1 m gages underestimated pit data by 39.5% on average over a set of 15 storm events. The smaller NRG and lower rim height placement provides the advantage of reduced deformation of the wind field (Kurtyka, 1953; Huff, 1955) but has the disadvantage of potential insplashing as noted previously. The NRG were placed in locations such that no vegetation would intercept an imaginary inverted 45° cone whose peak was placed at the gage orifice. Some sheltering of the NRG was inevitable as the desert brush forms a non-uniform canopy in the watershed. However, gage placement on a grid was closely adhered to provide an estimate of the distribution of rain falling on the watershed surface.

The recording gages and the vectopluviometers were installed at an elevation of 1 m. The recording gages employ a mechanical clock and record accumulated rainfall on a paper chart attached to a rotating drum. They were calibrated before installation and a calibration check was performed after the field measurements. A 6 h recording chart rotation with a theoretical resolution of 0.31 min $\pm$ 0.31 min was employed (Chery and Kagan, 1975). However, just as in chart depth resolution (Table 1), physical and data processing limitations such as ink line width, chart clip bulges, positioning of digitizing cross hairs, parallax and analog to digital converter resolution result less precise, post processing time scale resolutions of 0.6 min $\pm$ 1.2 min for the 6 h charts (Chery and Kagan, 1975).

Measurements of the non-recording gages were made after each storm when possible. TDR readings were performed every morning. Runoff measurements in Lucky Hills 104 were carried out by means of a supercritical flow flume equipped with a float-activated recorder (Smith et al., 1981).

3.2. Data used

Measurements were carried out from 4 July to 13 September 1990. Table 2 summarizes the data collected, basic statistics and quality assessments. A post-experiment calibration of the recording raingages eliminated data from four gages from further use (Table 2, footnote c). During the collection period, 18 independent rainfall events were measured with the recording raingages. The observed events covered a broad range of storm size and initial soil moisture conditions. Eight events produced significant runoff at the outlet of watershed LH-104. To help insure detection of small measured rainfall variations, a severely constrained selection of the events was made on the basis of measurement quality.

Eleven sets of measurements from the network of non-recording gages were used to determine the geostatistical characteristics of the storms. For these storms, the range

Table 2
Storm summary statistics and data quality ratings

Event	Date	Num. of indiv. events	Begin time[a]	Elapsed time to read NRG (h:mm)	Data quality code[b] A, V, R, T, D	Storm depth averages (mm)				Storm depth S.D. (mm)			
						VNRG[c]	RG[d]	TNRG[e]	CNRG[f]	VNRG	RG	TNRG	CNRG
2	07–05	1	1650	*	R D	*	33.3	*	*	*	0.2	*	*
3	07–11	1	2030	14:25	V, R, T, D	14.3	13.7	14.6	13.5	0.7	1.0	2.0	0.3
4	07–14A	1	1350	–		–	7.0	–	–	–	1.0	–	–
5	07–14B	1	1930	–	R D	–	7.6	–	–	–	0.2	–	–
6	07–15	1	0700	–	R D	–	20.0	–	–	–	0.4	–	–
4–6	07–14	3	1350	42:15	A, V	37.0	34.6	36.8	36.4	1.1	0.7	1.0	0.4
7	07–18	1	1640	01:15	T	2.6	1.9	2.7	2.6	0.2	0.3	0.3	0.1
8	07–19A	1	1915	–		–	–	–	–	–	–	–	–
9	07–19B	1	2230	–	R D	–	8.3	–	–	–	0.6	–	–
10	07–19C	1	2350	–		–	2.9	–	–	–	1.6	–	–
11	07–20	1	0050	–		–	4.1	–	–	–	0.3	–	–
8–11	07–19	4	1915	14:15	A, V	18.4	15.3	18.1	18.4	1.0	1.8	1.8	0.5
12	07–21	1	0450	53:10	V, R, T, D	7.2	7.9	7.5	7.0	0.7	0.8	1.4	0.2

1.1	0.3
2.7	0.7
0.3	0.1
3.2	0.3
–	–
–	–
–	–
0.8	0.1
2.8	0.6

work ($n = 48$ gages, 11
ges, 9 useable events); T,
ements with no known

Table 3
Comparison of storm statistics from 30 × 30 m and collocated vertical non-recording raingages

Event	Date	Storm depth				Coeff of var. percent (%)		Range (mm)			Number of VNRG outside 3σ Cheb. range
		Mean (mm)		S.D. (mm)							
		CNRG[a]	VNRG[b]	CNRG	VNRG	CNRG	VNRG	CNRG	VNRG	3σ Cheb. of CNRG[c]	
3	07–11	13.5	14.3	0.3	0.7	2.2	4.9	1.5	2.8	1.8	21
4–6	07–14	36.4	37.0	0.4	1.1	1.1	3.0	0.9	6.6	2.4	12
7	07–18	2.6	2.6	0.1	0.2	3.8	7.7	0.2	1.2	0.6	6
8–11	07–19	18.4	18.4	0.5	1.0	2.7	5.4	1.2	4.4	3.0	6
12	07–21	7.0	7.2	0.2	0.7	2.9	9.7	0.4	2.7	1.2	20
13	08–01	12.6	12.7	0.3	0.7	2.4	5.5	0.4	2.5	1.8	8
14	08–03	14.1	14.7	0.7	1.1	5.0	7.5	1.1	5.5	4.2	4
15	08–06	2.5	2.4	0.1	0.2	4.0	8.3	0.3	1.0	0.6	2
16	08–12	52.5	52.9	0.3	1.9	0.6	3.6	0.7	10.6	1.8	31
17–19	08–13	13.3	13.5	0.1	0.5	0.8	3.7	0.4	2.6	0.6	20
20	09–13	13.4	13.2	0.6	1.4	4.5	10.6	1.6	5.9	3.6	12

[a] Collocated vertical non-recording raingages within a 2 × 2 m area at location 606 ($n = 4$).
[b] Vertical non-recording raingages on 30 × 30 m grid ($n = 48$).
[c] 3σ Chebyshev range $= [m_{CNRG} + 3(SD_{CNRG})] - [m_{CNRG} - 3(SD_{CNRG})] = 2(3)(SD_{CNRG})$.

in mean rainfall from the 48 non-recording raingages was 2.4–52.9 mm with a mean of 9.9 mm (Table 3). This compares with the historical range and mean for recording raingage 83 of 0.25–72.6 mm and 5.5 mm, respectively for the period of July 1963 to December 1991 ($n = 1545$). The events are typical of air mass thunderstorms occurring in the monsoon season in Chihuahuan and Sornoran deserts of the southwestern United States and northern Mexico.

4. Analysis of results

4.1. Accuracy of non-recording gages

Variations owing to manufacture of the 16.2 cm^2 gages can be assumed negligible. However, the scale printed on the exterior of the gages proved to be too inaccurate to provide a good estimate of the amount of rainfall collected so readings were made by transferring the contents of the gage into one of two sizes of graduated cylinders whose estimated reading accuracy is presented in Table 1. Although this process induced a loss owing to the water adhering to the walls of the gages, it was a systematic error which has no effect on the assessment of spatial heterogeneities. Another source of error causing underestimation of the absolute rainfall depth from the non-recording raingages is evaporation during the elapsed time between the end of the storm and reading of the gages (Table 2, column 5). Gill (1960) noted daily evaporative losses up to 0.75 cm in (5.8 × 6.4 cm) for wedge-type gages for summer months at a Pennsylvania test site due to the relatively large surface area. However, Gill (1960) also noted that for a 3 cm diameter raingage daily evaporative losses were almost negligible for total rainfall amount less than 5 cm and approximately 1 mm for 10–15 cm rainfall amounts. For this study, in all but three of the 11 events, readings were taken the morning following a prior day's afternoon or evening thunderstorm. During the time between storm and reading it was assumed that evaporation from all the non-recording gages would be uniform owing to the relative homogeneity of temperature and radiation fields at the scale of the study area. It was also assumed that, over the typical time required to read the non-recording gages (~ 1.5 h), differential spatial changes in depth observations owing to evaporation would not be large enough to effect the assessment of spatial heterogeneities. However, comparisons between the recording and non-recording gages may be impacted.

Another source of error lies in the sampling method. The rainfall process is not homogenous. Microclimatic variations, vegetation sheltering, differential evaporation, and insplash conditions in the immediate neighborhood of the gages will induce a random measurement error. The four gages installed at site 606, within a 2 × 2 m, (Fig. 1(b)) provide an approximate estimate of these measurement variations. In Fig. 2, the coefficient of variation of these four measurements for each event is plotted as a function of the mean rainfall measurement. Also plotted are data from paired 6 and 24 h recording raingages ($n = 2$, separation less than 1 m) within the Walnut Gulch Watershed obtained by Osborn and Keppel (1966). Similar results were obtained by

Kalma et al. (1969) and Dreaver and Hutchinson (1974); namely, a poor relation between the mean rainfall and the standard deviation, and a decaying relation between the mean and the coefficient of variation. An estimate of the upper 95% confidence interval limit of the coefficient of variation for the latter two studies was drawn in Fig. 2 for comparison. These studies show that the measurement error can be quite considerable, falling from 10% for small rains (< 10 mm) to a possible asymptotic limit of 4% for heavy rains. The line at the bottom of Fig. 2 is an attempt to represent the part of the error which is due to the non-recording raingage reading inaccuracies described earlier. The broken line indicates a jump in measurement error at 15 mm of precipitation when a switch was made from using a 25 ml to a 100 ml graduated cylinder for measuring storm depths (see Table 1, footnotes a and b). For small storms (< 15 mm in which the 25 ml graduated cylinder was used, see Table 1), it can be conservatively concluded that one cannot expect an error of less than 5% for the non-recording raingages (also see Table 3).

In conclusion, the total error, conservatively estimated from the coefficient of variation, in point precipitation measurement for non-recording gages in the Lucky Hills experiment can be expected to be approximately 4–5% for storm event rainfall depths larger than 15 mm. For smaller events, it increases sharply, but no single measurement showed a variation greater than 6%. Approximately half the variation could be explained by reading error.

4.2. Accuracy of recording gages

For this type of gage, Chery and Kagan (1975) and Freimund (1992) note that the idealized resolution for a 1:1 depth scale to rain depth scale is 0.25 ± 0.25 mm

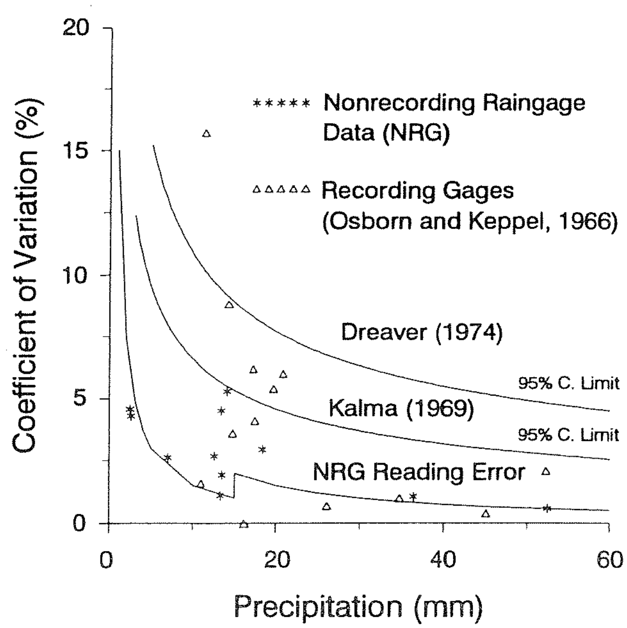

Fig. 2. Relative measurement error estimated from four closely spaced raingages at site 606.

(0.01 ± 0.01 in). In a postexperiment calibration check, the five selected gages showed no deviation in total depth over the calibration range greater than 0.3 mm. For total depth, the precision of data processing was estimated by digitizing the same chart five times for both the 5 July and 3 August events for raingage 602. The resulting standard deviation of the readings was approximately 0.07 mm and corresponds to the

4.4. Spatial variability of the rainfall field

To draw conclusions regarding the existence of spatial variability of rainfall over the small study catchment it must be established that observed variability is not an artifact of measurement error. To address this issue it was assumed that the collocated raingages provide an estimate of total measurement error for the non-recording gages. The mean and standard deviation from the collocated gages for each storm were then used with the Chebyshev inequality to estimate the probability that observations from the larger grid network of gages would fall within the range defined by the Chebyshev inequality (Benjamin and Cornell, 1970). For each event in Table 3, the four centrally collocated gages were used to compute a range defined by the Chebyshev inequality (see equation in footnote c of Table 3). The range was computed using 3σ bounds to provide a conservative range estimate resulting from measurement error. If the range of values observed from the grid network is larger then the computed Chebyshev range there is a relatively small probability $(1-(1/3^2)=1/9$ or less than 12%) that the total variation of rainfall over the grid (catchment) is the result of variation from measurement error.

Table 3 compares the statistics and ranges for the collocated and grid gages, the 3σ bound Chebyshev range, and the number of gages in the grid network with observations outside the Chebyshev range. In a storm by storm comparison, the range of grid gage observations exceeds the $3\ \sigma$ Chebyshev range computed from the collocated gages in all cases. In all but one event (3 August) the grid range exceeded 4σ bounds (< 7% probability that the total variation of rainfall over the grid is the result of variation from measurement error). In a second comparison, the total storm depth observation for each grid gage ($n = 48$) is examined to see if it lies within the $\pm 3\sigma$ interval about the mean of the collocated raingages. In this case, under the null hypothesis that measurement error is the only source of variation, the expected number of grid gages that will lie outside the Chebyshev range is less than $48/9 = 5.3$. The last column in Table 3 indicates that the null hypothesis can be rejected for nine of the 11 events. Although the Chebyshev inequality provides relatively imprecise bounds, other more standard significance tests such as the F-test are not applicable in the presence of drift (non-stationarity) with autocorrelation of errors (Delfiner, 1976).

A geostatistical analysis was performed to examine the data for these and other spatial characteristics. The theoretical basis and a description of the terms used below are described in Journel and Huijbregts (1978). The sample variogram $\gamma(h)$ for each storm was plotted for the isotropic case and for four directions in the anisotropic case (Englund and Sparks, 1990). The isotropic variogram of the individual events systematically exhibited a steep slope, while in the anisotropic cases, variograms from different directions behaved very differently. These observations, associated with the known properties of the storms at larger spatial scales (Osborn et al., 1979), suggested that the rainfall field was not stationary at the scale of the catchment.

An automatic structure identification algorithm (Delfiner, 1976) was used to detect drift in the data sets (stationarity is characterized by a zero-order drift; a first-order drift indicates a linear gradient in X and Y). The degree of the drift, and the drift

Table 4
Geostatistical analyses of the storms in Lucky Hills: characteristics of (1) drift, and (2) variograms of the residuals

Event	Date	Num. indiv. events	Degree of drift	Gradient of the drift			Ave. wind dir.[a] (Deg.)	Variogram of residuals		
				Mag. (mm per 100 m)	Mag. (%)[a]	Dir. (Deg.)		Nugget (mm^2)	Sill (mm^2)	Ratio (%)
3	07–11	1	1	0.88	6.13	79	120	0.160	0.225	71
4–6	07–14	3	0	–	–	–	–	0.500	1.600	31
7	07–18	1	1	0.28	10.88	135	129	0.015	0.030	50
8–11	07–19	4	1	1.36	7.39	144	–	0.400	0.550	73
12	07–21	1	1	1.04	14.50	287	251	0.110	0.140	79
13	08–01	1	1	1.11	8.72	83	114	0.110	0.110	100
14	08–03	1	1	1.58	10.76	145	32	0.375	0.620	60
15	08–06	1	0	–	–	–	–	0.020	0.022	91
16	08–12	1	1	2.48	4.70	310	250	1.200	2.000	60
17–19	08–13	3	1	0.58	4.32	149	–	0.125	0.140	89
20	09–13	1	1	1.63	12.33	11	–	0.500	0.900	56
Accumulations over longer periods										
07–21 to 08–12		5	0	–	–	–		2.000	3.000	67
07–11 to 09–13		19	0	–	–	–		6.000	10.000	60

[a] Magnitude of the gradient expressed in percent of the average rainfall depth.
[b] Rainfall depth-weighted average direction of wind vector.

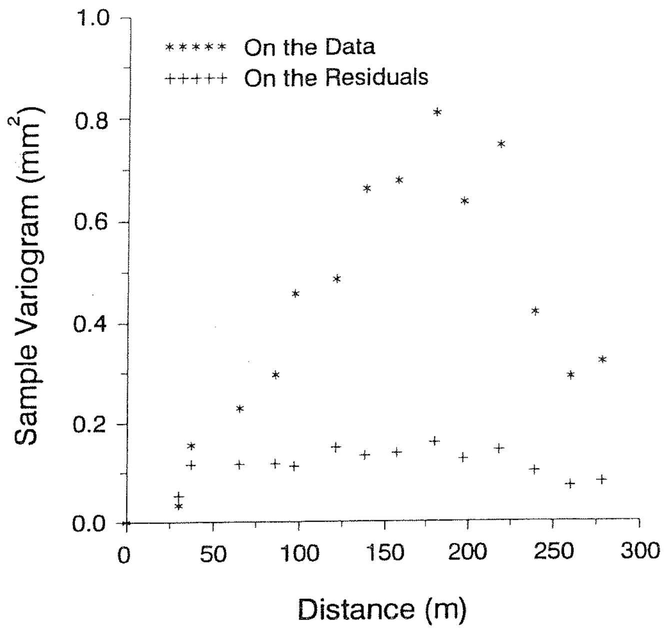

Fig. 3. Sample variograms for the data and residuals (after trend removal) for the storm of 21 July 1990.

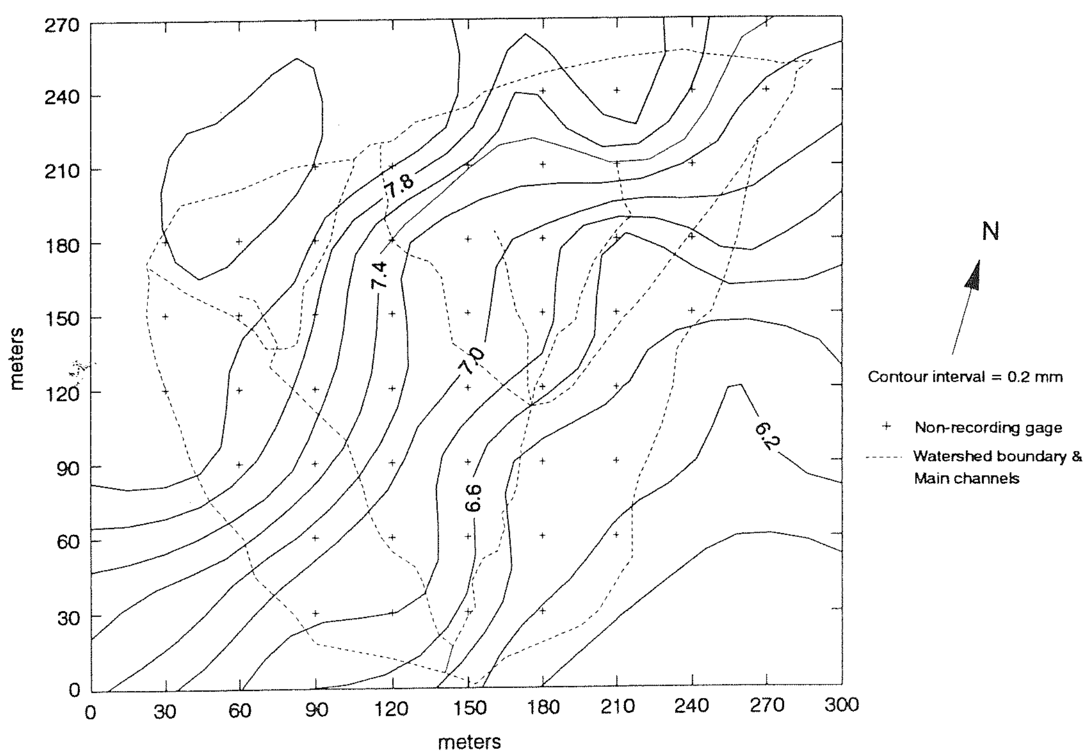

Fig. 4. Contour map of the rainfall depth for the storm of 21 July 1990 at Lucky Hills (interpolation by isotropic kriging).

gradient magnitude and direction are presented in Table 4 for each set of storm measurements. With the exception of the very small event measured on 6 August (1.8 mm), all the measurements involving individual events were found to have a first-order drift. The corresponding rainfall gradient, expressed in millimeters per 100 meters was calculated and included in the table. Sample variograms on the data and residuals (after first-order trend removal) for the 21 July storm are illustrated in Fig. 3. When measurements involved more than one rainfall event, the degree of the drift was either 0 or 1. Cumulative rainfall depths over longer periods (multiple events) showed a zero-order drift indicating compensating effects from event to event. Figs. 4 and 5 illustrate examples of storms exhibiting a relatively strong percentage gradient (21 July) and absolute gradient (12 August). On average, the rainfall gradient was found to be about 1.2 mm per 100 m. Hershfield (1969) reported that extreme rainfall gradients of 35 mm over a distance of 400 m (8.75 mm per 100 m) were observed in similar conditions from less dense gage networks.

Several reasons for the existence of drift at this small scale were explored. Attitudinal variation was explored first. No consistent relationship was found between total observed rainfall and elevation or the watershed elevation gradient and the drift gradient (Table 4). The magnitude and direction of the drift noted in this table is highly variable suggesting the drift may be related to factors which vary on a storm by storm basis such as localized wind. Assuming that the location of the air mass thunderstorms is relatively random in space the associated downdrafts can come from virtually any direction. To explore this the rainfall weighted average wind vector direction for each storm was also included in Table 4 where wind data were available.

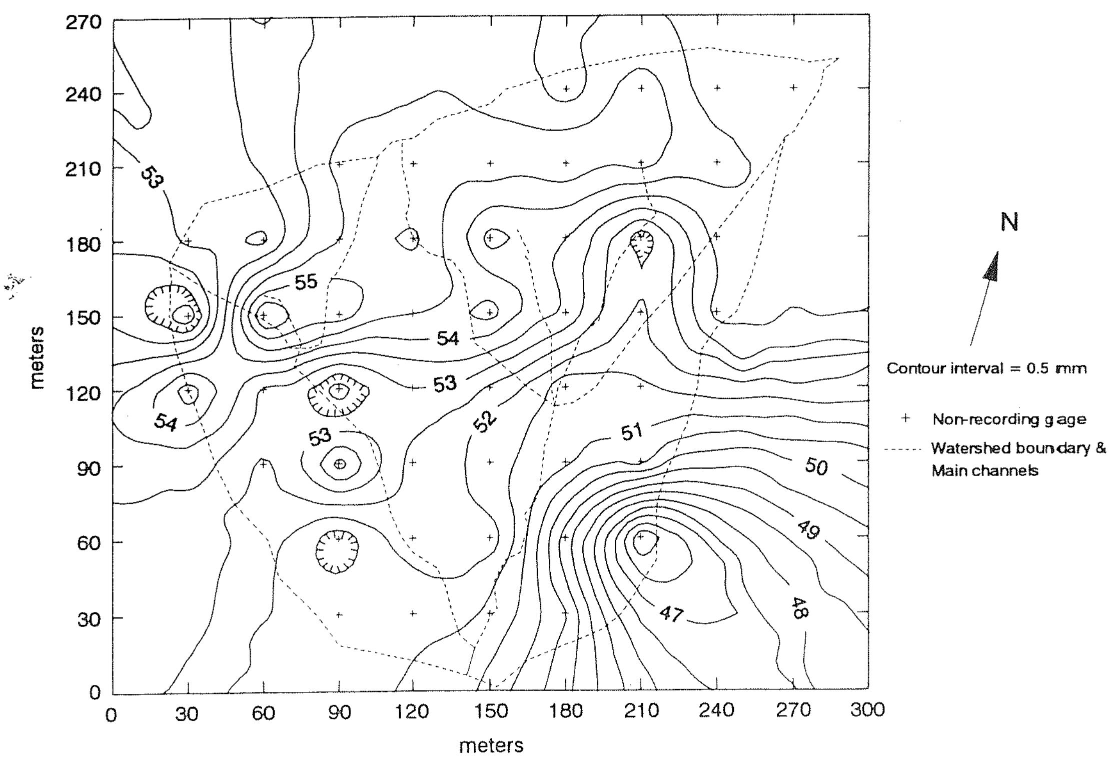

Fig. 5. Contour map of the rainfall depth for the storm of 12 August 1990 at Lucky Hills (interpolation by isotropic kriging).

Note the relatively good correspondence (correlation coefficient is 0.82) between drift gradient direction and the wind gradient direction suggesting that wind may be a primary cause of the drift.

For all cases except for the 14 July measurement (three storms combined), the nugget of the variogram of the residual (after drift removal) represents more than 50% of the sill (Table 4). This indicates that the drift explains a major part of the spatial correlation of the rainfall fields. The remaining variations are due largely to the measurement uncertainties described previously. In several cases, the variogram of the residuals could be represented by a pure nugget, suggesting a complete absence of spatial correlation in the residuals.

In conclusion, the determination of a first-order drift (trend) is sufficient to describe the spatial rainfall structure at the scale of the catchment in most cases. The rainfall gradients observed during this experiment induced variations from 4 to 14% of the mean rainfall depth over a distance of 100 m. In this small watershed the four collocated central raingages did provide a good estimate of the areal catchment precipitation computed from the arithmetic mean (Thauvin and Lebel, 1989) of the uniformly distributed raingages on the grid ($n = 48$). For all the events listed in Table 3 the maximum percentage difference in total storm rainfall from the central gages to the mean of the grided gages was −5.6% (event 3). With the existence of drift the error of areal average precipitation estimates resulting from the trend will be directly proportional to the gage distance from the center of the watershed if only one gage observation is available.

4.5. Spatial variability of time distribution of rainfall

Cross-correlation was used to extract information related to the spatial variations in rainfall intensity patterns (May and Julien, 1990). From the digitized rainfall charts of gages 601 and 609, the rainfall rates were computed for 1 min time intervals. The rainfall rate r at raingage location G_i, is a function of space and time ($r(G_i, t)$). The cross-correlation coefficient is given by

$$\rho_\tau(G_1, G_2) = \frac{1}{n-\tau}\sum_{k=1}^{n}\frac{[r(G_1, t_k)\delta_k - \bar{r}(G_1)][r(G_2, t_{k+\tau})\delta_k - \bar{r}(G_2)]}{S(G_1)S(G_2)} \tag{2}$$

where ρ_τ is the correlation coefficient at lag τ, t_k is the kth time step, δ_k is a rainfall indicator, n is the record length at zero lag, r is the mean of the time series and S is the biased estimate of the standard deviation. The rainfall indicator (δ_k) is designed to suppress long periods with no rainfall. It is equal to 0 when both rates are 0, and 1 in other cases. A distance of 170 m separates the two raingages used in this analysis. The correlograms for nine events were computed for these two stations. Fig. 6 illustrates the correlograms for the two events with the extremes of positive and negative time lag of peak correlation and for three additional events with smaller peak lags. The lag to peak correlation was examined in the context of rainfall weighted mean wind speed and direction and no consistent relationship between these quantities was found.

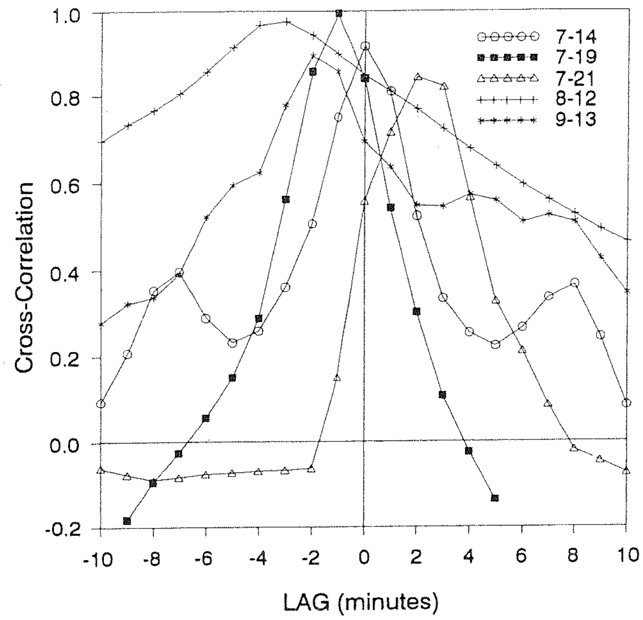

Fig. 6. Cross-correlation for 1 min time interval intensities at gages 601 and 609 in for five storms.

For the nine events the maximum correlation ranged from 0.83 to 0.99. Eight of the nine events had a time shift of the peak correlation of 2 min or less. This shift cannot be considered significant given the time resolution of individual gages (1 min) and the associated potential measurement error ±1.2 min (Chery and Kagan, 1975; Freimund, 1992). Within the measurement capabilities of the instruments used in this study, the assumption of a homogeneous time distribution of the rain over the catchment is reasonable at the scale of the study. This assumption implies that within the temporal accuracy of the recording raingages, the temporal distribution of rainfall intensities from each recording raingage can be considered equal if adjusted by linear scale factor to account for spatial variability in storm depth. Further investigation with more accurate techniques and dataloggers could be used to more carefully examine this conclusion. If the assumption of time homogeneity is acceptable, a single recording gage could accurately represent the temporal rainfall intensity distribution over a catchment of this size.

4.6. Wind analysis

During the experiment, measurements were performed to study the impact of wind on rainfall catch by both a direct and an indirect approach. In the direct approach, the tilted gages should give a measure of the rainfall catch by a hillslope considered homogeneous in slope and aspect. However, it should be reiterated that tilted gages do not solve the problem of local turbulence affecting rain catch.

To compare the results between collocated tilted and vertical gages, it was first assumed that with judicious placement of the tilted gages (see Fig. 1(b)), their rain catch represented subareas of similar size and that an arithmetic average of the results of individual gages accurately represents the mean areal rainfall over the catchment. The mean of the nine tilted gages was compared with the mean of the nine collocated vertical gages for each of the 11 rainfall events analyzed (Table 2). Assuming a normal distribution of the measurements and unknown variance, a t-test was performed to look for significant differences in mean areal rainfall depth. In no case did the two sets of measurements show a significant difference.

Two reasons can explain this result. First, the measurement variance, coupled with the small sample size ($n = 9$ vertical and tilted gage pairs) is important, and the expected variations are small given the small slopes of the watershed. Furthermore, because the measurements are averaged over the watershed area, a significant difference would be expected only when the rain comes from a direction parallel to the average slope of the watershed (NW–SE). In four cases (18 July, 1 and 13 August and 13 September), the main orientation of the rain was close to southeast and the mean from the tilted gages was systematically higher than the mean calculated with the vertical gages. On 12 August, the wind came from the northwest and the tilted gages have a lower average than the vertical gages. These results indicate that, owing to the relatively gentle topography of the watershed, the small sample size, and the measurement errors, differences between tilted and vertical gages are hardly perceptible.

The indirect approach to the problem of rainfall inclination employs vectopluviometer data which measures the vertical and four cardinal horizontal components of rain 1 m above the ground from which the storm average rainfall inclination and aspect can be computed (Hamilton, 1954). The results from the three vectopluviometers (location in Fig. 1(b)) were compared with wind data from the climate station. The climate station anemometer to record wind vector magnitude (m s^{-1}) and direction (degrees) was installed at an elevation of 4 m, on the top of the hill forming the northern edge of the catchment (Fig. 1(b)). The vectopluviometer gives a measure of the average wind vector magnitude and direction expressed in terms of rainfall inclination. For each climate station recording period (15 or 20 min), rainfall inclination (b) and direction (z_b) were then simulated from wind measurements for each rain storm where wind and vectopluviometer data were available. The results of the simulation are compared with the data from the vectopluviometers and a linear regression with wind calculated by the two methods resulted in the coefficients reported in Table 5.

The correlation between the anemometer and the gages decreases significantly

Table 5
Coefficients of the linear regression of wind calculated by vectopluviometers compared with the data from the anemometer

Gage	Constant (m s^{-1})	Slope	R^2
697	0.40	0.979	0.895
698	0.73	0.912	0.792
699	0.73	0.500	0.573

when the distance from gage to anemometer increases. Furthermore, the slope of the regression line (m) sharply decreases from gage 697 to gage 699 (Fig. 1(b)). This probably is due not only to the increasing distance from the anemometer but also to wind exposure of the gage. The change in gage location results in a decrease of average calculated wind speed from 3.1 m s^{-1} at gage 697 to 1.9 m s^{-1} at gage 699. Average inclination also decreases from 22° at gage 697 to 14° at gage 699.

The altering of wind direction is indicated in Fig. 7 where the mean wind direction is represented by an arrow at each of the three gage locations. It appears that at the outlet of the catchment, the wind was systematically forced to the direction of the main channel (NW–SE). For the two other gages, it is much more variable. When the average wind direction was within 90° of the average slope of the catchment (NW–SE) the paired vertical and tilted raingages also both tended to catch more rainfall in the lower portion of the catchment.

These results suggest that even at the scale of this experiment, the use of a correction factor calculated from a single vectopluviometer should be made with caution due to the apparent inhomogeneous rainfall inclination field which resulted in variable rain catch across the catchment. Impacts on hydrologic rainfall measurements are mitigated in this catchment due to the relatively smooth topography as indicated by the tilted versus vertical raingage measurements at paired locations. In

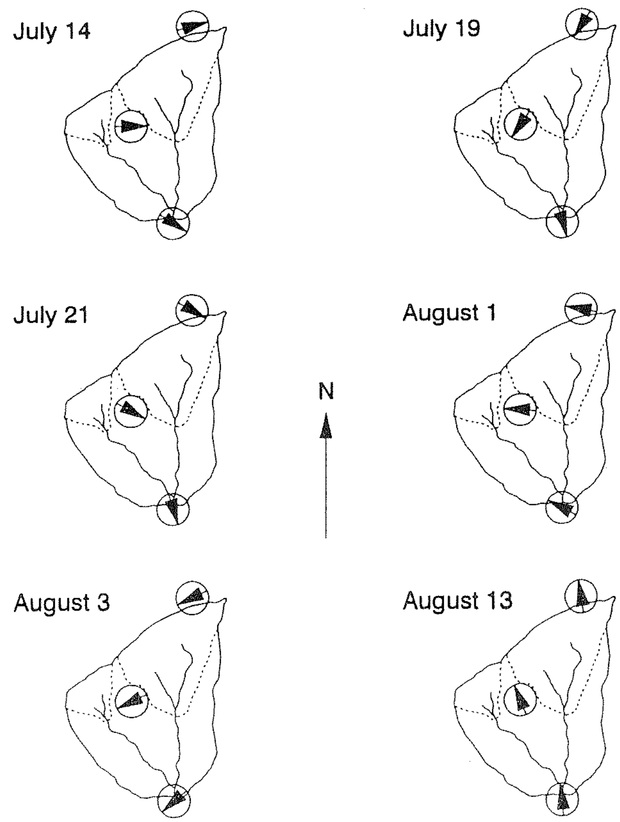

Fig. 7. Average wind direction calculated from vectopluviometer data at three locations in Lucky Hills 104.

the companion paper by (Faurès et al., 1995, this issue) the impacts of wind induced rainfall inclination and catch differences from vertical and tilted raingages on runoff computations are explored.

5. Conclusions

A dense raingage network of recording and non-recording gages was established in a 4.4 ha catchment in southeastern Arizona to test the assumption of small catchment spatial rainfall homogeneity. A variety of factors were considered in the study to test this assumption. They include: estimates of rainfall data accuracy for recording and non-recording gages, systematic measurement error, data reduction impacts, quantification of spatial variability, spatial variability of the time distributions of rainfall, and impacts of wind.

The study lead to the following conclusions.

(1) The total error in point rainfall measurement for non-recording gages was estimated to be 4–5% of total storm depth for storms larger than 15 mm with increasing errors for smaller storms, conservatively estimated at 10% for storms in the 2–5 mm range. For the recording raingages the potential variation from total measurement error around the observed value is ± 0.8 mm for total depth with a time measurement error of ± 1.2 min (Freimund, 1992). For total depth, using a 0.8 mm error, percentage errors are comparable to those from the non-recording raingages (approximately 5% for a 15 mm storm with increasing error for smaller storms and decreasing error for storms larger than 15 mm). There was no significant difference in total event rainfall depth between larger and higher recording raingages and non-recording gages.

(2) Using bounds computed from the Chebyshev inequality it was found that the range of observed variation over the watershed was greater than the variation that would result from total measurement error. In addition, geostatistical analysis indicated the presence of first order drift with corresponding rainfall gradients ranging from 0.28 to 2.48 mm per 100 m with an average of 1.2 mm per 100 m. These gradients represent a 4–14% variation of the mean rainfall depth over a 100 m distance indicating that raingage location is particularly important if only one gage is available. This suggests that the typical uniform rainfall assumption is invalid at the 5 ha scale in regions where convective thunderstorm rainfall is significant. This has important implications as testing and validation of process-based hydrologic models are often conducted on small research watersheds using the spatially uniform rainfall assumption (single raingage). The impacts of observed rainfall variability on runoff modeling is presented in a companion paper (Faurès et al., 1995, this issue).

(3) The instrumentation utilized in the study did not detect a significant difference in the temporal rainfall intensity distribution for eight of nine events examined. If the assumption of time homogeneity is acceptable a single recording gage can accurately represent the rainfall intensity distribution over a catchment of this size and inexpensive non-recording raingages could be used to quantify the spatial rainfall

distribution. This type of measurement network could circumvent the use of a large number of recording raingages as this option is typically uneconomical.

(4) Direct and indirect methods of analysis of the impact of wind on the spatial rainfall distribution indicated that for the relatively gentle topography of the study catchment and inherent measurement errors there is little perceptible difference between catch in tilted and vertical gages. However, this does not imply that wind impacts are not important, particularly in regions of more pronounced topographic variation. Vectopluviometer data indicated that the assumption of a homogeneous rainfall inclination field is not valid, even at the scale of this experiment.

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