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Radar precipitation for winter hydrological modelling

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Abstract Interpolated precipitation gauge measurements and weather radar snowfall estimates were used as input to a physically-based hydrological model (WATCLASS). The gauge measurements were corrected for wind undercatch according to WMO standards. Post-processing of the radar data was undertaken to consider underestimation due to the use of winter radar coefficients for liquid precipitation. Streamflow was simulated for the Upper Grand River basin in central southwestern Ontario, Canada for the five winters from 1993 to 1997. For each year, except 1995, the radar data provided precipitation estimates that were better, in terms of simulated runoff volumes, than those provided by the gauge data. Along with runoff volumes, peak streamflows were more closely estimated from radar precipitation than gauge precipitation. Gauge estimates consistently yielded lower than observed peak streamflows.

Key words precipitation gauges; snow hydrological modelling; weather radar; winter hydrology

INTRODUCTION

For hydrological modelling, weather radar can provide the spatial distribution of rainfall, whereas the gridding of point gauge rainfall produces questionable areal estimates, especially where gauge data are sparse. For colder climates, gauge snowfall estimates require correction for wind undercatch and other errors (Goodison *et al.*, 1998) and the use of weather radar for snowfall estimation has focused primarily on short time intervals, such as hourly and sub-hourly, to single storm events lasting up to two days. Improvements in radar estimates have emphasized the use of ground-based measurements as truth.

Numerous *Z–R* relationships have been developed for snowfall (Fujiyoshi *et al.* (1990) provides a good summary of previous *Z–R* relationship determinations). Most are based on individual storm events and involve comparisons to precipitation gauges. The analysis by Sekhon & Srivastava (1970) developed the *Z–R* relationship that is widely used for the cold weather season precipitation estimation. Recently the advanced capabilities of weather radar have been explored for complex snowfall identifications. For example, Ryzhkov & Zrnic (1998) used polarimetric radar to distinguish between rain and snow, and Matrosov (1998) showed that the difference between the reflectivities at the dual wavelengths yielded a snowflake size estimate that could be combined with the longer wavelength reflectivity to approximate a snowfall rate.

The use of weather radar to estimate snowfall across watersheds for modelling purposes has been limited to work by Houck *et al.* (1995) who recommended using weather radar snowfall estimates as input to a GIS hydrological analysis, and Fassnacht *et al.* (1998) who used weather radar for modelling the winter season on the Grand River basin in southern Ontario.

This paper investigates the use of weather radar *vs* gridded gauge data as the precipitation inputs to a hydrological model. Peak streamflows and cumulative runoff volumes are compared for five winters across a 3520 km^2 watershed in southern Ontario, Canada.

STUDY AREA

The Upper Grand River watershed is located in central southwestern Ontario and has a climate that is moist with cold winters (a Koppen Dbf climate, as per Gough *et al.*, 2002). Synoptically, it is located near the moving boundary between continental polar air originating in northern Canada, maritime polar air from the Pacific modified by the Rockies, maritime tropical air which forms over the Gulf of Mexico, and continental tropical air from the south central area of the North American continent (Gough *et al.*, 2002). The polar air flows in from the northwest and is dominant in the winter. Winter temperatures are warmed due to the Great Lakes, and snow belts extend from Lake Huron into the northern portions of the basin.

The Meteorological Service of Canada (MSC) radar in King City, north of Toronto, covers the basin. The installation currently operates in a Doppler mode, but the imagery used in this paper are 2 × 2 km hourly precipitation accumulation maps derived from a composite of C-band (5.2 cm) conventional scans of 10 min Constant Altitude Plan Position Indicator images. For the warm season, the *Z–R* coefficients developed for southern Ontario by Richards & Crozier (1983) are used, and for the cold season, the Sekhon & Srivastava (1970) *Z–R* coefficients are used.

The watershed is mostly covered by crops and low vegetation (59%), with some regions of wetland (18%), mixed deciduous–coniferous forest (14%), small bare (8%) and impervious (1%) areas. The terrain in this portion of southern Ontario is glacial material, composed primarily of clayey till. Streamflow was monitored at eight hydrometric stations across the study basin for the study period, i.e. the winters of 1993–1997.

METHODOLOGY

Hydrological modelling

The WATCLASS physically-based distributed hydrological model was used to simulate streamflows. This is a linkage of the University of Waterloo WATFLOOD hydrological model and the Environment Canada CLASS land surface scheme (Soulis *et al.*, 2000). The vertical water and energy budgets near the land surface are computed using CLASS, while the lateral water budget and streamflow routing is performed using WATFLOOD. The Upper Grand River basin was modelled using 10 × 10 km grid elements, for which the slope and flow direction have been determined by Tao & Kouwen (1989).

Gauge undercatch correction

Precipitation was measured using Nipher-shielded gauges. Gauge undercatch was estimated using wind speed data (Goodison *et al.*, 1998) and snowfall was corrected to the DFIR reference. The corrected gauge data were further corrected to the bush setting (Yang *et al.*, 1993). The gauge data were corrected for wetting loss by adding 0.15 mm to each precipitation event.

Meteorological data gridding

The WATCLASS model requires air temperature, precipitation, barometric pressure, specific humidity, wind speed, shortwave and longwave radiation data. Gauge data were gridded using the inverse squared distance interpolation (Tao & Kouwen, 1989). Longwave radiation data were computed from the air temperature, specific humidity, and an estimate of the cloud cover. The daytime cloud cover estimates were derived from a ratio of measured to computed shortwave radiation (Fassnacht *et al.*, 2001a).

Radar data adjustment

For the study area, Fassnacht *et al.* (2001b) identified a local scaling issue, that caused overestimation of precipitation during periods of strong anomalous propagation. Reprocessing of the data to the lowest possible increment minimized the forced over-scaling problem.

Radar reflectivity from melted or partially melted hydrometeors yield precipitation underestimates when snowfall Z–R coefficients are used for rainfall, freezing rain, etc. The probability of snow *vs* air temperature curve derived by Auer (1974) was used to estimate the quantity of mixed precipitation and the proportion of rain was adjusted based on the rain Z–R coefficients (Fassnacht *et al.*, 2001b).

RESULTS

Areal estimation of snowfall is a problem related to gauge density across a watershed and to the technique used to grid the point data. The use of weather radar can overcome the gridding problem. For example, the 10 km resolution gridded gauge precipitation, using the inverse squared distance approach, and the 10 km radar precipitation for 26 January 1996 from 12:00 to 13:00 h EST are presented in Fig. 1(a) and (b). From the various gauges across central southwestern Ontario, precipitation was only observed at the north central gauge, around which the gridded precipitation is

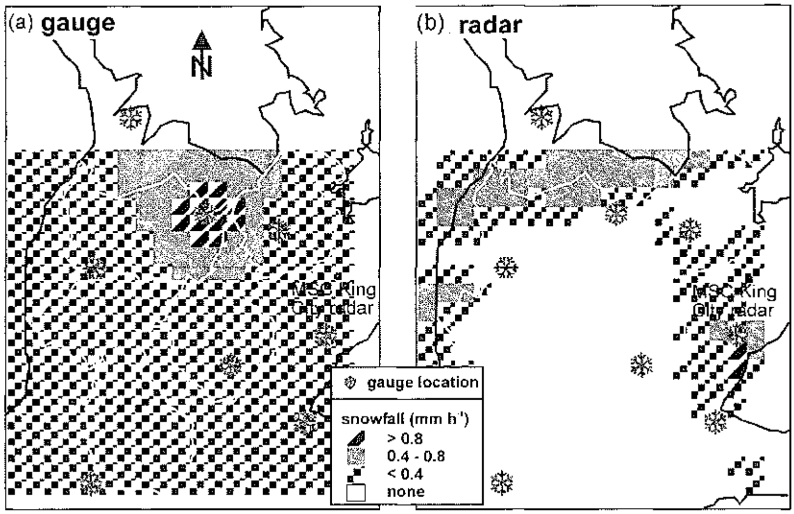

Fig. 1 Hourly precipitation at a 10 km resolution across the five central southwestern study basins on 26 January 1996 at 13:00 EST derived from: (a) gridded gauge data, and (b) radar data.

concentrated. No precipitation was observed at this location according to the radar. Precipitation across the remainder of the domain is a result of gridding (Fig. 1(a)). From the radar estimates, precipitation was only observed over the Toronto station in the eastern portion of the domain, just below the MSC King City radar (Fig. 1(b)). However, the radar precipitation near Toronto has often been observed to be urban clutter.

On 28 January 1996 from 02:00 to 03:00 h EST, no precipitation was observed at any of the gauges. Thus, the gridding of the gauge data yielded no precipitation (Fig. 2(a)). The left-middle gauge was not recording and the right-middle gauge did not measure precipitation. However, radar observations indicated the presence of precipitation, but not near the recording gauges (Fig. 2(b)). The discrepancy between the spatial gauge and radar precipitation is due primarily to the gridding of the point data.

Streamflows were simulated at eight stations within the basin for the winters of 1993–1997 using the gridded gauge and adjusted radar precipitation data. These model results were compared to the observed streamflows that were obtained from the HYDAT CD-ROM (Environment Canada, 1997).

A double mass curve of cumulative modelled *vs* observed streamflow at the mouth of the basin for the winter of 1993 illustrated that the adjusted radar provided a better precipitation estimation for streamflow modelling than did the gridded corrected gauge data (Fig. 3). This is consistent for both peak flow and runoff volume at all hydrometric stations for 1993, as illustrated by the residual plots Fig. 4(a) and (b), respectively. The plots illustrate the difference between modelled and observed peak flows or runoff volumes using radar precipitation *vs* using gauge precipitation at the eight stations for the five study years. A better radar than gauge precipitation estimate for modelling yields a data point closer to the x-axis (the shaded area) and a better gauge estimate yields a data point closer to the y-axis (unshaded area).

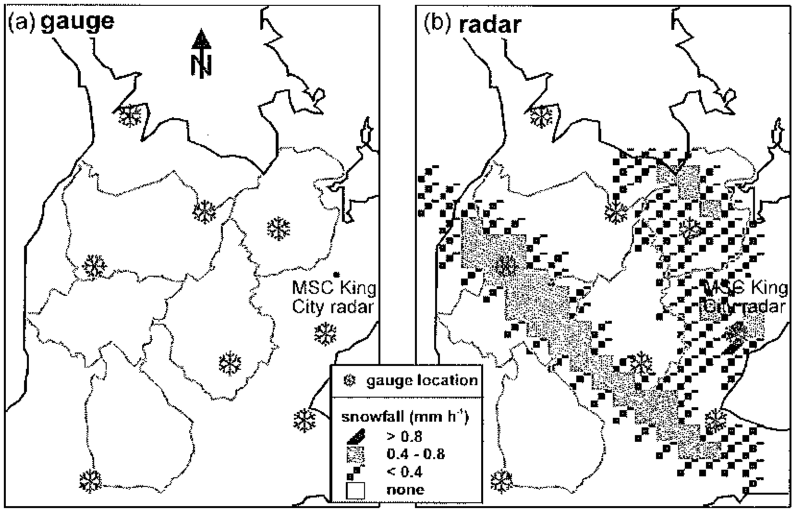

Fig. 2 Hourly precipitation on 28 January 1996 at 03:00 EST derived from (a) gridded gauge data, and (b) radar data.

There is little annual net difference in peak runoff for 1994, 1995, and 1997, typically with varying extents of over- and underestimation from modelling (Fig. 4(a)). Peak flows are more appropriately modelled with gridded gauge data than with radar data for 1996. Comparatively, radar was slightly better overall in terms of absolute error. However, 80% of the modelled peak flows using gauge precipitation were underestimated with an average peak flow underestimation of 23.1 $m^3 s^{-1}$. Using the radar precipitation, the modelled peak flows were almost evenly distributed in number between underestimates (47.5%) and overestimates (52.5%) with an average overestimation of 14.1 $m^3 s^{-1}$.

The peak streamflows from radar data were overestimated for 1995, and removal of the 1995 results yielded an average peak flow overestimation of 1.66 $m^3 s^{-1}$ for the radar precipitation versus an underestimation of 36.7 $m^3 s^{-1}$ for the gauge precipitation.

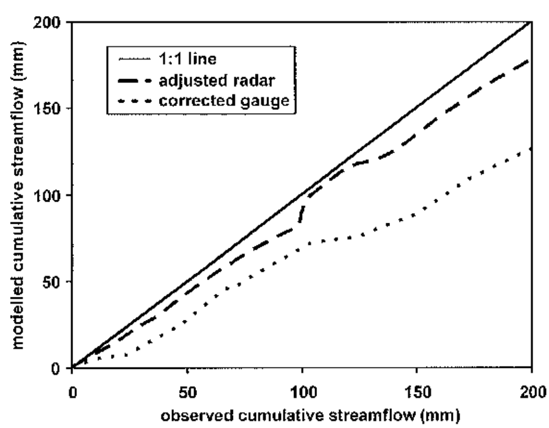

Fig. 3 1993 winter cumulative modelled and observed streamflow at the mouth of the basin.

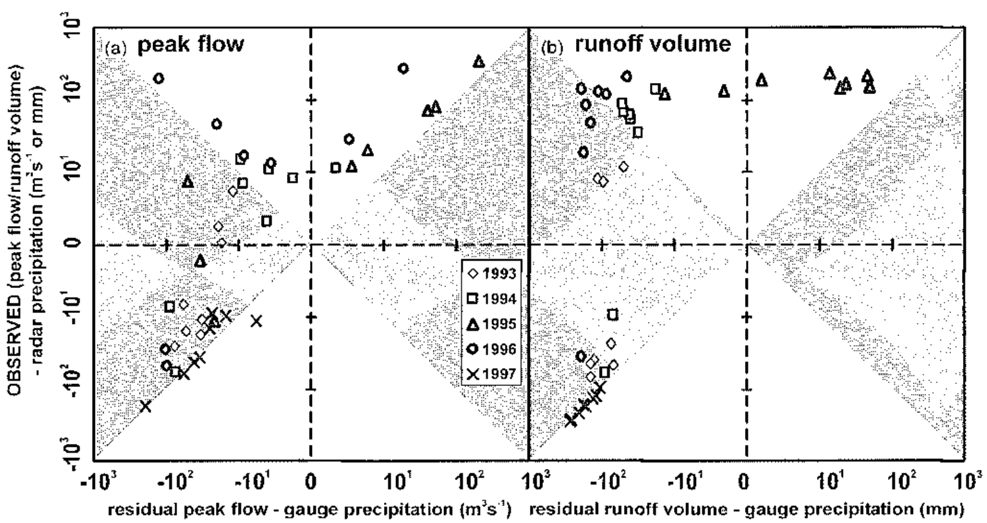

Fig. 4 Residual of modelled minus observed (a) peak flow and (b) runoff volume from snowmelt streamflow for radar precipitation *vs* gauge precipitation. Each symbol denotes one winter's peak flow at one hydrometric station.

The discrepancy for 1995 was more obvious from the simulated runoff volumes using radar precipitation (Fig. 4(b)). Whereas the 1996 peak flows were better estimated from gauge precipitation, the 1996 runoff volumes were better estimated from radar precipitation. The results from 1994 and 1997 were similar. However, for all years except 1995, the gauge precipitation yielded underestimates of simulated runoff volume. The gauge precipitation yielded on average 91 mm less simulated runoff and the radar yielded 20.1 mm more simulated runoff. Ignoring the 1995 data yielded an average underestimation of 41.7% from the gauge precipitation and an underestimation of 0.3% from the radar precipitation.

DISCUSSION

The radar precipitation yielded better simulated peak streamflows and runoff volumes for four of the five study years. The gridded gauge precipitation yielded underestimates for most sub-watersheds for most years. These differences occurred in part due to the gridding of the precipitation data. An inverse weighted distance scheme was used to grid the gauge precipitation data, as per Tao & Kouwen (1989). However, they modelled summer storm events and there are more rainfall gauges than snowfall gauges across this and most study areas. The use of another distance-type interpolation scheme, such as kriging, should not improve the gridded gauge underestimation. Similarly, the use of hypsometric interpolation would be limited by the lack of gauge data, as well as the limited topographical variation across the region.

The gauge data have been corrected to consider undercatch due to wind. Without this correction, the gridded gauge data yielded larger underestimates of streamflow. The radar data have been corrected due to a local scaling issue, which reduced much of the radar overestimation, and resulted in underestimation of streamflow. The data were adjusted to consider underestimation by using winter *Z–R* coefficients for liquid precipitation. This adjustment improved the radar precipitation estimation.

The radar precipitation overestimation for 1995 was a problem, as illustrated by Fassnacht *et al.* (2001b). The 1997 precipitation was underestimated by the radar. Similar patterns were observed during the 1995 and 1997 winters for the state of Michigan from the National Weather Service NEXRAD system (Hollingsworth, personal communication, 1999). These variations may be due to differences in synoptic storm types. These years must be further investigated to remove seasonal systematic biases, when and where they exist.

Due to systematic seasonal differences in radar precipitation estimation and the lack of station data, it is recommended that a combination of the gauge and radar data be investigated for estimating winter precipitation. This combination of gauge and radar data has been performed for rainfall with different degrees of success by Wilson (1970), Brandes (1975) and others.

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