An Extreme Rainfall/Runoff Event in Arctic Alaska

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ABSTRACT

Rainfall-generated floods in the Arctic are rare and seldom documented. The authors were fortunate in July 1999 to monitor such a flood on the Upper Kuparuk River in response to a 50-h duration rainfall event that produced a watershed average in excess of 80 mm. Atmospheric conditions prevailed that allowed moist air to move northward over areas of little or no vertical relief from the North Pacific Ocean to the Arctic Ocean. Cyclogenesis occurred along the quasi-stationary front separating maritime and continental air masses along the arctic coast. This low-pressure system propagated southward (inland) over the 142-km^2 headwater basin of the Kuparuk River in the northern foothills of the Brooks Range; a treeless area underlain by continuous permafrost. This research catchment was instrumented with a stream gauging station, two major and six minor meteorological stations, for a total of eight shielded rain gauges. The peak instantaneous flow was estimated at 100 m^3 s^{-1} and was about 3 times greater than any previously measured flood peak. Historically in the Arctic, annual peak floods occur following snowmelt when the snowpack that has accumulated for 8–9 months typically melts in 7–14 days. The shallow active layer, that surficial layer that freezes and thaws each year over the continuous permafrost, has limited subsurface storage when only thawed to a depth of 40 cm (at the time of the flood). Typically for this area, the ratio of runoff volume to snowmelt volume is near 0.67 or greater and the ratio for cumulative summer runoff and rainfall averages around 0.5 or greater. For the storm discussed here the runoff ratio was 0.73. These high runoff ratios are due to the role of permafrost limiting the potential subsurface storage and the steep slopes of this headwater basin.

1. Introduction

We hypothesize that while a majority of the annual floods in the Arctic are snowmelt-generated, most floods of record will be rainfall-generated with the exception that as the watershed size increases the likelihood of the measured flood of record resulting from rainfall decreases. Compared with more temperate regions, most high-latitude stream gauging stations have shorter record lengths; so the probability of observing floods of high return periods (or low probability) is reduced. It is clear that the likelihood of major rainfall-generated floods at the high latitudes is relatively low; this just means that snowmelt-generated floods will dominate short time series of maximum annual floods. Watershed size is also important, as is the orientation of the watershed. The likelihood of a large frontal system delivering significant rainfall to a high percentage of large arctic basins like the Mackenzie River or the Lena River is very slight. However, during the long winter season, the likelihood that the entire basin will be covered with

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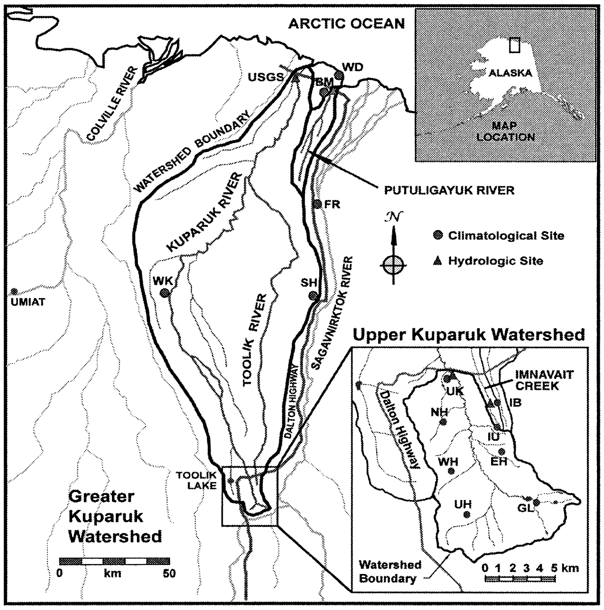

FIG. 1. A map of the nested watersheds being studied on the North Slope of Alaska, including hydrological and meteorological station locations. Details of the Upper Kuparuk River are shown in the lower right-hand corner.

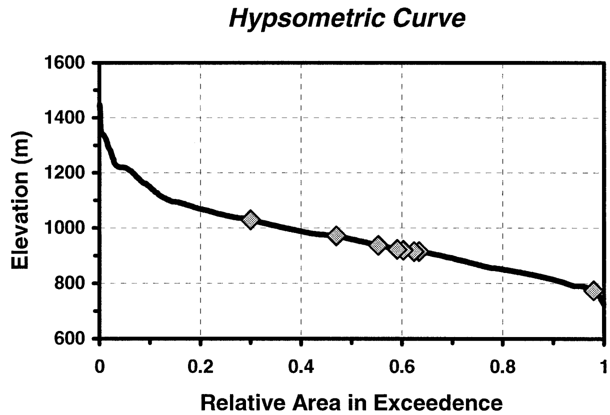

FIG. 2. Hypsometric curve for the Upper Kuparuk River. The markers on the curve represent the elevations of each rain gauge; the lowest mark is also the stream gauging site.

snow is essentially 100% at winter's end. For north-draining rivers, the melt usually starts in the lower latitudes and moves northward concurrent with the routing of the northward flowing runoff. This results in higher peak flows than when the ablation starts near the watershed outlet and proceeds up into higher elevations of the basin.

The rate of snowmelt is limited by the amount of excess energy available at the surface (Kane et al. 1997), while rainfall rate is controlled by many factors that include the amount of moisture that physically can be held in the atmosphere (related to temperature), cyclogenesis, topography, and others. The energy fluxes that are important during snowmelt are long- and shortwave radiation and latent and sensible heat fluxes. Advected energy from areas that are already snow-free can also be important. For a relatively small drainage (142 km^2) on the North Slope of Alaska, the Upper Kuparuk River, we monitored a large rainfall event during July 1999. This precipitation event generated a runoff response that was approximately 3 times larger than any previously measured discharge during the 1993–2001 period.

What physical conditions need to prevail that would result in significant rainfall on the North Slope of Alaska? Low air temperature, mountainous topography, and an extensive ice cover over the Arctic Ocean all combine to reduce the amount of precipitable water in the atmosphere over the Arctic. First, the typically colder atmospheric temperatures limit the amount of water vapor potentially contained in the atmospheric column. This dictates that air masses laden with potential precipitation must originate in warmer environments. Second, moist air masses from the south often lose much of their precipitable water when passing over mountain ranges as they progress northward; a route obviously without mountains would result in higher moisture levels being transported to the high latitudes. Finally, the extensive ice cover on the Arctic Ocean sharply reduces the surface flux of water into the atmosphere for most months of the year.

The runoff response of arctic watersheds is significantly impacted by the presence of permafrost (subsurface storage and surface storage), topography (energy gradients and surface storage), the antecedent hydrologic processes (precipitation, evapotranspiration, and soil moisture) and the rainfall or snowmelt duration and intensity. The net result is that permafrost enhances the fraction of both rainfall and snowmelt (when the active layer is also seasonally frozen) that leaves a basin as runoff, particularly in high-gradient watersheds. Kane et al. (1989) examined several storms where there was no response of runoff from rainfall events as large as 12–15 mm (no precipitation in the 5 days preceding the rain event). It is assumed that this water went primarily into active-layer storage.

2. Study area and instrumentation

The Upper Kuparuk River drains the northern foothills of the Brooks Range, flowing northward towards the Arctic Ocean (Fig. 1). This watershed and some additional nested watersheds have been the focus of a study to enhance our understanding of high-latitude hydrology over the past few years (McNamara et al. 1998; Kane et al. 2000). The main basin length of the Upper Kuparuk catchment is 16 km and the channel length is 25 km; the drainage area is 142 km^2. The stream gauging site is at an elevation of 698 m with a maximum watershed elevation of 1464 m (Fig. 2). Alpine vegetation communities are prevalent at higher elevations and tussock sedge tundra at the lower elevations. Dwarf willows and birch up to 1 m in height occupy riparian areas along the stream channels and water tracks.

Deep subpermafrost groundwater is precluded because permafrost is continuous under the catchment with maximum depths of about 250 m. Glacial till from numerous glaciations is mantled with organic soils of varying depth from little or none on the ridges to 50 cm or more in the valley bottoms. The active layer at summer's end is approximately 40–50 cm deep in poorly drained sites and up to 100 cm in well-drained sites (Hinzman et al. 1998); an average value when the flood occurred in mid-July of 40 cm would be typical. This constitutes the only potential subsurface storage. There are two small lakes in the headwaters with an area of a couple of hectares that represent the major surface storage.

One major [Upper Kuparuk (UK); Fig. 1] and five minor meteorological sites [North Headwater (NH), West Headwater (WH), Upper Headwater (UH), Green Cabin Lake (GL), and East Headwater (EH)] are located within the Upper Kuparuk watershed; in addition, there is a major meteorological site just to the east about 1 km from the watershed boundary in Imnavait Creek (IB) and a lone precipitation gauge at the head of Imnavait Creek (IU) and located just outside the watershed. Spatially these sites are well distributed (Fig. 1), but there are no precipitation gauges in the highest 30% of the basin (Fig. 2). At the major sites, profiles of air temperature, relative humidity, and wind speed are measured over a height of 10 m. Other sensors include wind direction, incoming and reflected solar radiation, incoming and emitted longwave radiation, net radiation, snow depth, rainfall precipitation, and soil and snowpack temperatures. The minor meteorological sites have instrumentation at one height only to measure wind speed, air temperature, and rainfall precipitation. Imnavait Creek has been instrumented since 1985, while the Upper Kuparuk watershed has been instrumented since 1993.

Standard U. S. National Weather Service (NWS) rain gauges (diameter ~20 cm, tipping bucket) equipped with Alter shields are used to measure rainfall. Measurement errors during the storm are limited to wetting losses that would be small compared to total storm precipitation (Yang et al. 1998). At the outlet of each basin that we study, stage–discharge rating curves are developed. During snowmelt period, the Upper Kuparuk watershed is gauged twice daily to coincide with low and high flows. The rationale for this procedure is that the stage–discharge relationship changes daily until all of the ice is eroded/melted out of the channel. Just prior to snowmelt, the channel is usually fully occupied by aufeis and snow.

3. Storm dynamics

For several days preceding the flood event that peaked on 17 July, Alaska was dominated by a persistent ridge oriented diagonally NW–SE across the state, bringing seasonably warm temperatures and generally clear skies to most of interior Alaska and the Arctic Slope (area north of the Brooks Range in Alaska). By 14 July [all dates and times are coordinated universal time (UTC)] the upper ridge had amplified, with lows forming over the central Aleutian Islands and British Columbia, Canada. This pattern provided the large-scale circulation that transported relatively warm, moist air northward from lower latitudes. It is interesting to note that many of the heavy precipitation events that have occurred across Alaska during the warm season have involved a similar meridional flux of warm and moist air of high equivalent potential temperature [K. Gilkey, NWS Fairbanks Weather Station Forecast Office (WSFO) 2000, personal communication]. During the next 2 days, the low over the Aleutians drifted north into the Bering Sea. Several weak short-wave troughs spun off from this low, moving northeastward up and around the ridge as the long-wave pattern shifted eastward. During this period a number of weak surface lows formed and dissipated over the NE arctic coast. This is a persistent region of frontogenesis during the brief arctic warm season as atmospheric surface temperatures over the land rise in response to essentially continuous solar forcing, while offshore the boundary layer is dominated by marine stratocumulus and sea surface temperatures are only a few degrees above freezing. Baroclinicity of 10°–15°C over a distance of less than 50 km is not unusual here in favorable conditions.

By 1200 UTC 16 July [0300 Alaska standard time (AST) 16 July] a more persistent surface low was centered over the coast along the Alaskan–Canadian border, with a trough extending over the eastern section of the Arctic Slope in Alaska. Precipitation was occurring along the trough, and a dewpoint temperature of 12°C was observed at Umiat, one the few regularly reporting stations on the Arctic Slope. During this time several of the meteorological stations also recorded a period of precipitation. Circulation around the low was cyclonic but quite weak, with the front essentially locked onto the coastline. In the next 24 h, the flow over western Alaska became more northerly at upper levels as the low over the Bering Sea moved eastward and a strengthening low over the North Chukchi Sea deepened to become the dominant pressure feature to the west. By 1200 UTC 17 July (0300 AST 17 July), the cyclonic flow around the deepening but stationary surface low (Fig. 3a) had strengthened significantly. The strong baroclinic zone that had been stationary along the coast moved southward as a vigorous and active cold front. This surface cyclogenesis was reinforced by a rapidly developing low at upper levels that had propagated northeast from the Bering Sea as a short-wave feature. Figure 3b shows contours of the 500-hPa geopotential height and as well, winds and temperature (shaded) at the 700-hPa pressure level. Strong cold advection is apparent over central Alaska at 700 hPa. While the coldest temperatures at this level were well south of the Arctic

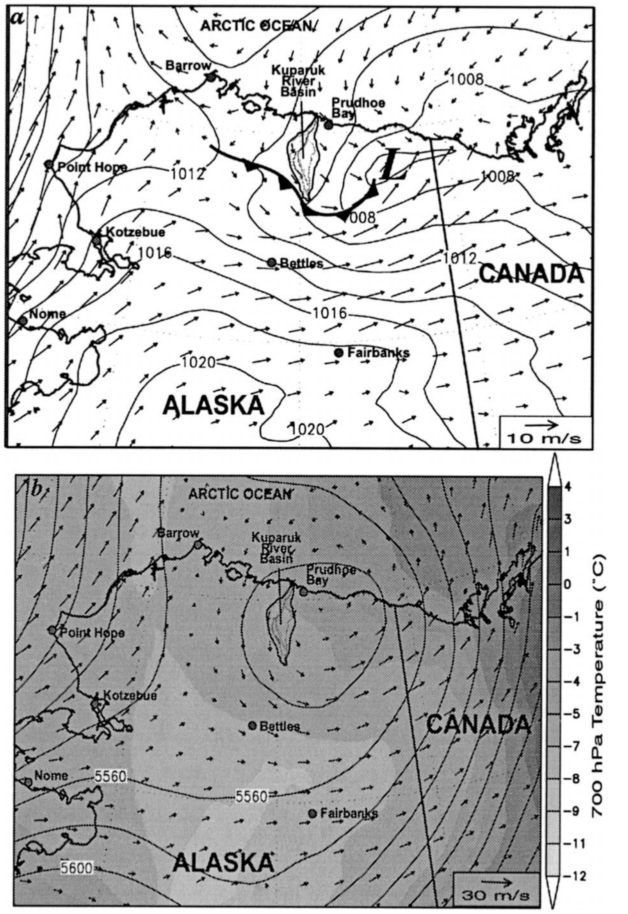

FIG. 3. (a) Surface winds (m s^{-1}) and mean sea level pressure (mb contours) for 1200 UTC 17 Jul. (b) Height of 500-hPa surface (m, contoured at 20-m increments), 700-hPa winds (vectors), and 700-hPa temperature (°C, shading).

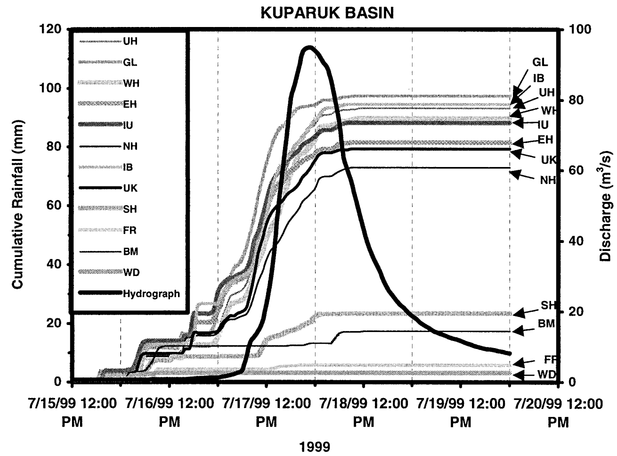

FIG. 4. Cumulative rainfall at eight headwater rain gauges and four other gauges between the Upper Kuparuk River catchment and the coast. The dark line represents the hourly mean discharge for the storm; note that the hourly mean discharge is slightly less at the time of the peak than the instantaneous peak of 100 m^3 s^{-1}.

Slope, advection over the Arctic Slope was sufficient to drop 700 hPa temperatures by as much as 5°C in the preceding 24 h.

These concurrent events combined to create an environment conducive to strong convective precipitation. The convergence and lifting in the boundary layer along the advancing cold front provided a focused forcing mechanism for boundary layer ascent. The strong advective cooling above the boundary layer in the 850–700-mb layer conditionally destabilized the lower troposphere, providing a good convective environment for parcels in saturated ascent along the frontal boundary. This was embedded within a broader mesoscale region of vertical ascent associated with the positive vorticity advection in the midtroposphere (Holton 1992), a factor often associated with convective storms (Barnes and Newton 1985). This combination of factors taken together (Schaefer et al. 1985) suggests that vigorous convection was quite likely over the Upper Kuparuk basin during the heaviest precipitation period.

This potent combination of concurrent events provided the setting for the heavy rainfall directly responsible for the flood event. The dynamical forcing aloft provided an environment of large-scale positive vertical motion (lifting) while the strong frontal forcing at the surface provided a focused mechanism for convection. The bulk of precipitation from this storm fell in the 12-h period starting 1200 UTC 17 July (0300 AST 17 July). During this time the cold temperatures at 700 hPa destabilized the atmospheric column thermodynamically, creating a conditionally unstable convective environment.

By 0000 UTC 18 July (1500 AST 17 July) the stacked low pressure system had moved north and eastward into Canada as high pressure was building back into the region from the west. The surface front dissipated as it moved southward into the higher terrain of the Brooks Range and was no longer apparent in the observations. While precipitation continued to fall over the meteorological network for the next 12 h, it was generally less intense and more typical of a postfrontal stratiform nature.

4. Rainfall input and runoff response of watershed

On 16 July, the first day of significant precipitation, approximately 20 mm of rainfall fell over the entire Upper Kuparuk River catchment (Fig. 4). Gauges farther north and at lower elevations had considerably less precipitation; for example, Sagwon Hills (SH) on the southern edge of the coastal plain and Betty Met (BM) near the coast had about 9 and 12 mm, respectively, on the same day.

While the four northern gauges (SH, BM, WD, and FR) collected very little additional precipitation on 17 July, all eight of the gauges in the Upper Kuparuk catchment collected another 50–70 mm (Fig. 4). The highest

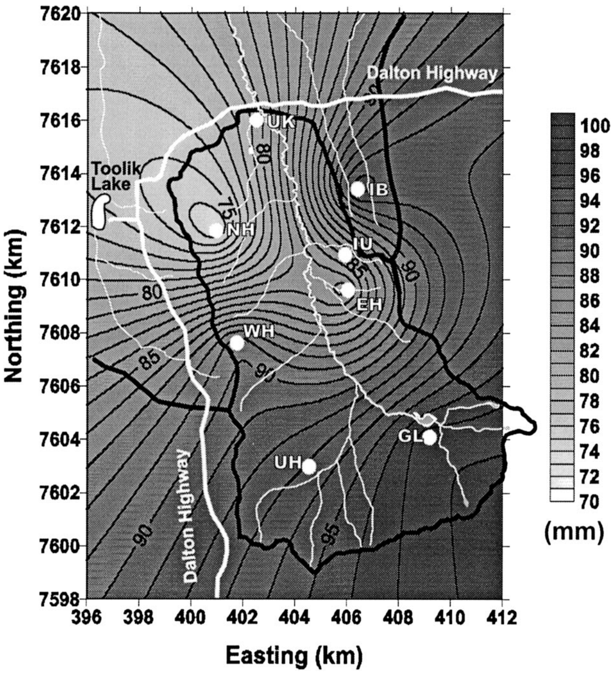

FIG. 5. Distributed rainfall over the Upper Kuparuk River catchment.

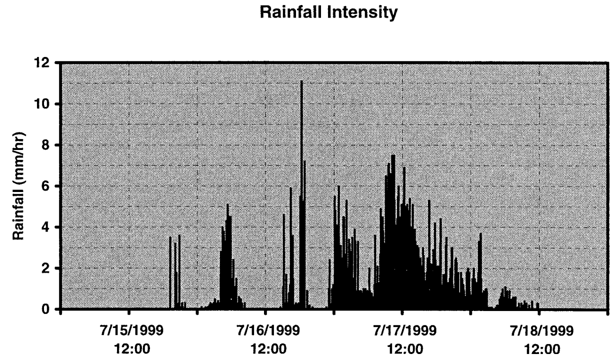

FIG. 6. Hourly rainfall intensity for the eight headwater rain gauges shows the periodic nature of the precipitation over the basin, with the most persistent precipitation at midday 17 Jul and the highest rainfall intensity on the 16th.

rainfall intensities occurred from early morning and lasted until late afternoon on the 17th. A few more mm of rainfall fell on the morning of 18 July before the storm ended. In Fig. 5, the spatial distribution of the total rainfall, kriged over the basin, for the entire storm is shown. The total rainfall ranged from 73 to 98 mm, with more precipitation falling in the headwaters and at higher elevations as would be expected.

Snow can fall on any day in this part of the Arctic; however, during this storm the air temperature was fairly warm (5°C) and uniform across the watershed and no snow on the ground was observed.

An examination of hourly rainfall intensities (Fig. 6) showed that there were six episodes of precipitation followed by little or no rainfall for 2–7 h. Through the fifth episode of rainfall the volume increased over the watershed. The lone exception to this trend was during the third episode when the rainfall was spatially sporadic. However, it was during this episode when the highest rainfall intensity of 11.1 mm h^{-1} occurred at the Imnavait Creek gauge (IB). The highest rainfall intensities at all of the other sites ranged from 5.3 to 7.5 mm h^{-1}. For many parts of the world these rates are not impressive, but they are for the Arctic.

5. Indirect methods of peak flow estimation

The storm in July 1999 produced a stage that far exceeded (by 0.87 m) the highest measured stage on our prestorm rating curve for that time at gauging site, although it did not disrupt the continuous stage measurement during the flood. The flood peak discharge (Q, $m^3\ s^{-1}$) was estimated indirectly using Manning's equation (Benson and Dalrymple 1967), also referred to as slope/area method:

$$Q = (1/n)R^{2/3}S^{1/2}A, \tag{1}$$

where n is Manning's roughness coefficient, R is the hydraulic radius in m, S is the water surface slope, and A is the cross-sectional area of the flow in m^2. Manning's n must be estimated and all of the other variables (R, A, and S) are measured in the field after the storm from high water marks. The channel was divided into five sections to account for different n values of the surface material (Fig. 7). Equation (1) was solved for each section, then the flows in each section were summed to get the total peak flow. Manning's equation was developed to be used for uniform flow where the energy, water surface, and bed slopes are all the same.

High-water marks were identified at the cross-section site soon after the flood receded (Fig. 7). The water surface slope was obtained by surveying the elevations of high water marks along a straight reach bounding the cross section. Although the cross section changed during the flood (Fig. 7), R and A from the preflood cross

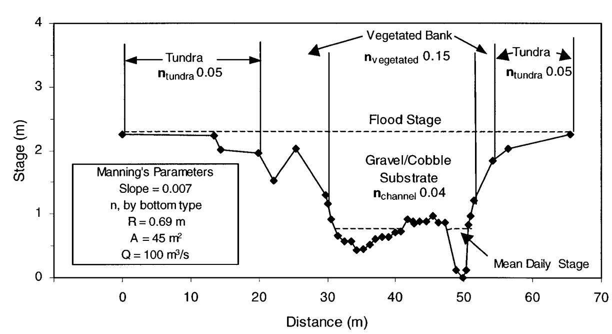

FIG. 7. Cross section on the Upper Kuparuk River where the indirect estimate of the flood was determined, this site also coincides with the stream gauging site.

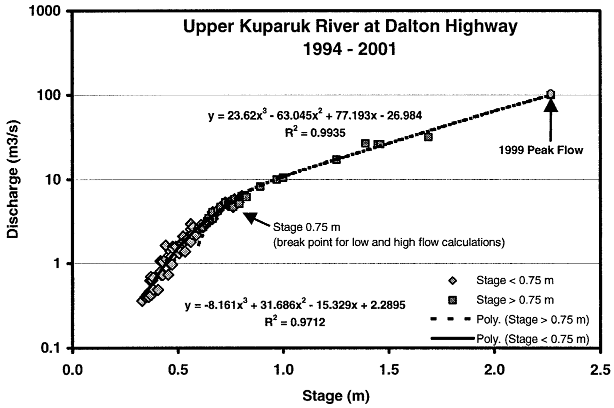

FIG. 8. Stage vs discharge curve for the Upper Kuparuk River (1994–2001).

section were used to solve Eq. (1). Manning's roughness coefficients for the vegetated sections and vegetation-free channel were selected by examining calculated *n* values for in-channel flows at Imnavait Creek and Upper Kuparuk River and also consulting Manning's roughness values found in several publications (Arcement and Schneider 1984, 1989; Barnes 1967). Because *Q* is so sensitive to *n*, it is very important that we select realistic values.

A continuous flood hydrograph (Fig. 4) was constructed using the estimated peak flow from Manning's equation, the measured stage record during the storm and a slightly adjusted stage-discharge relationship (Fig. 8) that included the new data point from this flood. A comparison of the stage–discharge relationships before and after the storm showed that there was no change.

The peak flow using Eq. (1) was estimated at 100 m^3 s^{-1}; the storm hydrograph is shown in Fig. 4. The flow started to increase modestly on 16 July from a prestorm flow of 0.65 m^3 s^{-1}; however, on 17 July the flow increased very rapidly and in fact peaked near the end of the day when rainfall intensities decreased. This peak is almost 3 times greater than any measured or estimated peak for this catchment; the previous high occurred during a snowmelt runoff event in 1997 and was measured near 34 m^3 s^{-1}. The maximum measured stage during the flood at the stilling well where the rating curve exists was 2.27 m. The rating curve we used prior to this flood would have predicted a flow of 105 m^3 s^{-1} for the maximum flood stage. The highest gauged flow during the summer is 27 m^3 s^{-1} at a stage of 1.4 m. We have had higher flows during snowmelt (e.g., 34 m^3 s^{-1}, 1997), but the stage–discharge relationship is meaningless during snowmelt when the channel has erodable snow and ice in the bottom with water flowing over it. Although it is not advisable to use a rating curve to forecast flows for stages beyond the range of measured data, the relative agreement between the Manning-derived estimate (100 m^3 s^{-1}) and an estimate from our prestorm stage–discharge relationship (105 m^3 s^{-1}) gives some credence to the flood peak estimate.

6. Measured and estimated floods in region

There are three nearby drainages in addition to the Upper Kuparuk River catchment that are presently gauged (Fig. 1): Imnavait Creek (2.2 km^2), entire Kuparuk River (8140 km^2), and Putuligayuk River (471 km^2). Details of these drainages are given in Kane et al. (2000). Both the Upper Kuparuk and Imnavait catchments are totally contained in the foothills. The Putuligayuk catchment is totally confined to the low-gradient coastal plain while the entire Kuparuk basin has its headwaters in the foothills but flows across the coastal plain before emptying into the Arctic Ocean (about 38% of the basin area is below 150-m elevation).

An examination of plots of the maximum snowmelt and rainfall floods for each year and for each catchment (Fig. 9) shows that most annual floods are snowmelt. Over a 17-yr period (1985–2001), there was only one maximum annual flood due to rainfall (1999) for Imnavait Creek. For a 9-yr period (1993–2001), there were three maximum annual floods for the Upper Kuparuk River due to rainfall (1993, 1995, and 1999). The Putuligayuk River was gauged from 1970 to 1995 (1980 and 1981 missing, 1987–95 estimates from crest gauge) by the U. S. Geological Survey (USGS) and by us continually from 1999–2001; all annual peak floods have been from snowmelt generation. In fact, there is little or no runoff response to summer precipitation on the Putuligayuk River because of the surface storage that develops over the summer when evapotranspiration exceeds precipitation (Bowling et al. 2003). For the entire Kuparuk River (1971–2001, gauged by the USGS), only the 1992 annual runoff peak is from rainfall. It should be noted that the snowmelt runoff that year was also quite low.

We have carried out flood frequency estimation using the Log Pearson III distribution for each of these watersheds. Because the physical mechanisms of runoff generation are entirely different for rain and snowmelt we have followed the recommendations of Waylen and Woo (1982). In this case we have performed flood frequency analysis for each process separately; this requires that we identify both the rainfall and snowmelt runoff peak each year.

For all the snowmelt flood sets, the coefficient of skewness is negative (Table 1); this means that for the higher-return periods the estimated floods are not increasing very rapidly. However, for the rainfall-predicted floods the coefficient of skewness is positive for those two watersheds in the foothills, meaning that at higher-return periods the flood estimates are increasing more rapidly than those for snowmelt. This is not surprising; an examination of a histogram of snowmelt floods (Fig. 9) for Imnavait Creek and the Upper Kuparuk River shows that the maximum flood is only 2 to

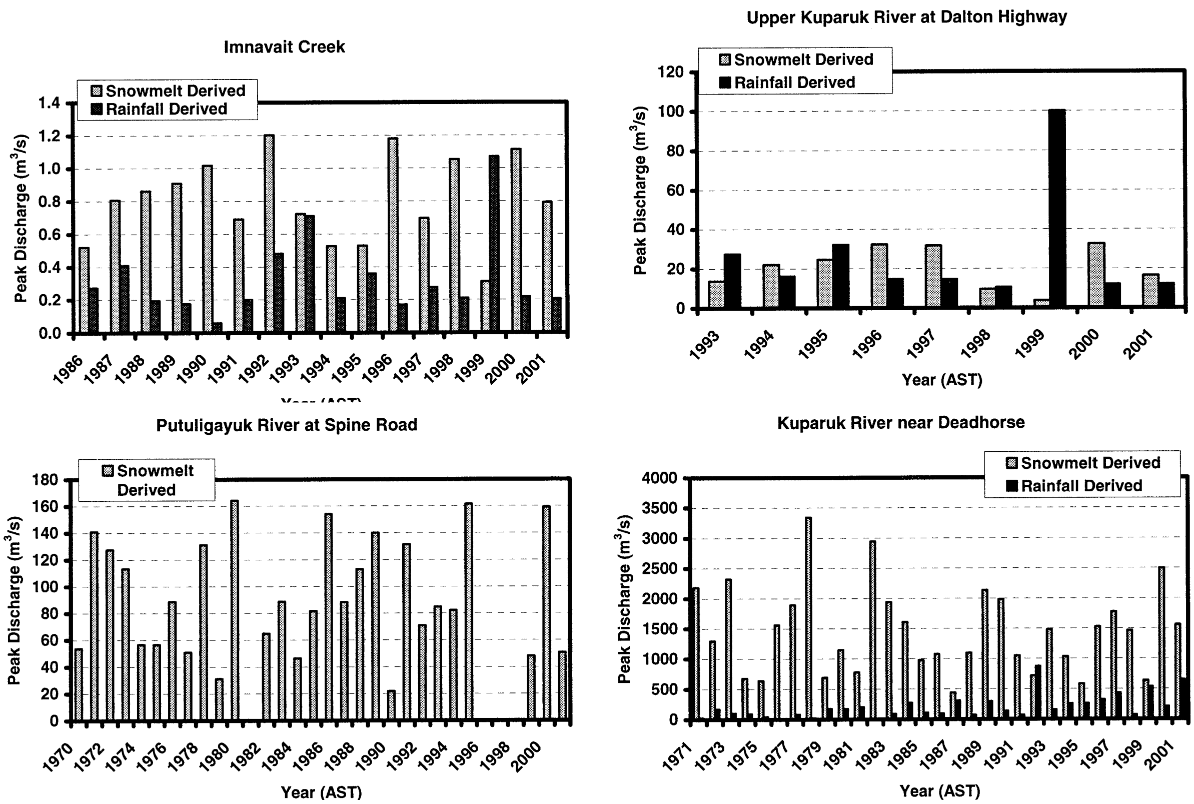

FIG. 9. Annual peak runoff events for both snow and rain generated for period of records for nested watersheds (no summer flows shown for the Putuligayuk River because of low flows).

3 times greater than the minimum snowmelt flood. For the rainfall-generated floods the differences between the maximum and minimum vary by a factor of 10 or more.

7. Discussion

From a rainfall perspective, potential summer floods in the Arctic would seem small and not very interesting. But, arctic watersheds have several distinguishing hydrologic traits. First from a structural viewpoint, they are underlain by continuous permafrost and are generally void of trees. The presence of continuous permafrost precludes the typical groundwater system, thereby severally limiting subsurface storage and baseflow. This lack of subsurface storage inflates the runoff response at storm and seasonal timescales of high-latitude watersheds relative to more temperate watersheds. McNamara et al. (1998) described the typical runoff response of arctic watersheds. In summary, snowmelt runoff dominates the annual runoff contribution and for individual summer storms watersheds exhibit a fast initial runoff response, long lag times between hyetograph and hydrograph centroids, an extended recession, and a high runoff/precipitation ratio. The rapid response is due partially to the existence of water tracks, small drainages found in permafrost environments that efficiently convey water downslope perpendicular to the elevation contours (McNamara et al. 1999). Numerous explanations have been proposed to explain extended recessions in arctic streams including slow release from highly absorptive organic soils (Likes 1966), low evapotranspiration rates (Dingman 1973), and the absence of appreciable baseflows (McNamara et al. 1998). The long lag time between hyetograph and hydrograph centroids is simply a result of the extended recession shifting the hydrograph centroid to the right.

TABLE 1. Comparison of the coefficient of skewness (of the logs) for snow and rainfall-generated floods.

Watershed	Coef of skewness snow	Coef of skewness rain
Imnavait Creek	−0.9113	0.0983
Upper Kuparuk River	−1.3541	1.6986
Kuparuk River	−0.1906	−0.6010
Putuligayuk River	−0.6303	*

* There are no significant rainfall-generated runoff events for this low-gradient basin.

For a watershed the size of the Upper Kuparuk River, the record peak runoff will come from rainfall because the potential rate of precipitation is greater for rainfall than for snowmelt. For this storm, the runoff response is 0.73. The reasons for the high runoff responses are discussed in several papers (Kane et al. 1998; Lilly et al. 1998; McNamara et al. 1998), but the limited subsurface storage because of the discontinuous permafrost is the primary reason. Once the subsurface and surface storage are completely replenished, there are no other alternatives for delaying runoff over short periods of time such as this storm.

McNamara et al. (1998) examined a range of storm responses that produced simple hydrographs; maximum response ratios for Imnavait Creek and Upper Kuparuk River were both 0.61 (1994 and 1995). Kane et al. (1998) reported on the hydrologic responses to individual storms (1996 and 1997) and got ratios of 0.16 to 0.55 for Imnavait Creek and 0.21 to 0.42 for the Upper Kuparuk River. During snowmelt when the active layer is completely frozen, runoff ratios exceeding the 0.67 of this flood are common. McNamara et al. (1997) found that the active layer played a minimal role in the runoff processes during snowmelt as most of the water leaving the basin came directly from the snowpack (most of the soil water is frozen at this time). During the summer, most of the water leaving the basin during storms was older water that had resided in the active layer from the previous summer or longer and was displaced by newer water. During snowmelt, transpiration is minimal; intense snowmelt takes place generally over a 7–14-day period (approaching summer solstice) and the snow water equivalent ranges from 80 to 180 mm (Kane et al. 1997).

House and Pearthree (1995) and Quick (1991) discuss the difficulties of making flood estimates by indirect methods described earlier. In a case study at Bronco Creek in Arizona, House and Pearthree concluded that initial peak flood estimates should be reduced by a factor of more than 2 (2080 $m^3\ s^{-1}$ down to 750–850 $m^3\ s^{-1}$). The main difficulties in obtaining good flood estimates (Costa 1987; Jarrett 1987; Baker 1987; Quick 1991) are determining accurate estimates of Manning's *n*, delineating the channel cross section at the time of the peak flow (channel erosion or deposition during storm), partitioning the energy used for water and sediment transport, and determining the water and energy slopes. The upper historical envelope curves of peak flood estimates for the United States have increased with time; this is partially due to expanded opportunities (more stations and longer records) to observe and estimate these floods. Some of the increase could be due, however, to overestimates of the floods when applying Manning's equation.

In many past floods where estimates were made by indirect measurements, there was little complementary hydrologic data. In our case, we have substantial precipitation data, continuous stage record, preflood channel cross sections and prior flow history with a stage–discharge relationship. We feel that our estimate is relatively good considering some of the potential errors that could misdirect our estimates. Since we used preflood cross sections this should reduce our flood estimate. Our greatest error is probably in not accounting for the energy used to transport sediment. We have very little data on sediment transport, but observations of channel changes are indicative of substantial transport during the storm.

We believe that it is clear that floods of record in the foothills of northern Alaska (return periods greater than 100 yr) will be rainfall generated. For the data collected to date on the two rivers (Kuparuk and Putuligayuk) that cross or drain the coastal plain it is not clear that this is the case. These two rivers have the longest gauging record but there are no rainfall floods for the Putuligayuk (28 yr) and only in 1 yr (a low snowmelt runoff year) for the Kuparuk did the rainfall peak exceed the snowmelt peak (31 yr). From the hypsometric curve for the Kuparuk River (Kane et al. 2000), it can be seen that 38% of the basin is below 150-m elevation, while the Putuligayuk basin is totally contained on the coastal plain (does not even extend to the base of the foothills) and has a maximum elevation of 109 m.

McNamara et al. (1998) found that the runoff contribution from the upper basins was much greater than coastal areas in contributing to the storm flow. Both Rovansek et al. (1996) and Mendez et al. (1998) found for a small coastal wetland that runoff essentially ceased during the summer months. The main reason for the lack of runoff on the coastal plain is the available storage that develops over the summer as evapotranspiration exceeds precipitation in the extensive lakes, ponds, and wetlands. Bowling et al. (2003) demonstrated that there was a substantial reduction in surface water over the summer for the Putuligayuk basin and also that the drainage network in the basin becomes fragmented.

8. Conclusions

Although the data reported here is of a short duration, a general understanding of the hydrologic and meteorologic processes of this nested grouping of arctic basins is evolving. It is comprehensible that for the headwater basins in the foothills, rainfall-generated runoff events will produce flood magnitudes that far exceed those generated during snowmelt. From the precipitation data collected to date, it is also obvious that the foothills receive significantly greater amounts of precipitation than the lower coastal areas and the higher the foothills the greater the annual precipitation received. This combined with the high-gradient topography and continuous permafrost ensures that the runoff response will be significant and rapid.

The hydrology of arctic regions is poorly understood spatially and temporally although there are pockets of research where considerable progress is being made. Present interest in the hydrology of high latitudes is motivated partially by interest in climate and related change; and how does this change impact high-latitude hydrology and what are the Arctic Ocean and atmospheric implications and feedbacks? However, it will take a long time to capture gradual hydrologic change induced by climate. It is more likely that extreme events, like the flood described here, will prove to be more important than gradual change. These extreme events can take many forms; rain on snow, midwinter melt, heavier or lighter than normal snowpacks, drought, and extreme rainfall. Since many hydrologic and meteorologic records in high latitudes are of short duration and often not complementary, coupled with a sparse network poorly distributed, it is often difficult to ascertain hydrologic impacts of climate change or even capture through field measurements storms like the one described here. For these reasons, it is important that index watersheds that are adequately instrumented be established and continued in the Arctic.

Obviously, for large arctic watersheds like the Lena, Ob, Yenisei, and Mackenzie that drain several climatic zones the major annual runoff event will continue to be from snow as these basins annually have essentially 100% snow cover at the end of winter and rainfall-intense low pressure systems can only produce rain on a fraction of the basin area at a given time. But other factors like the topographic gradients, watershed orientation, and combinations of hydrologic processes can result in unique behavior. For example, the 471 km^2 Putuligayuk River catchment has demonstrated no significant response to summer precipitation (Fig. 9) for the 28 yr of observations (not continuous). The two main factors for this lack of response are the low topographic gradient and the surface storage deficit (evapotranspiration exceeds precipitation over the summer). Several factors control the snowmelt runoff response, but arctic catchments will under almost any combination of factors produce a significant snowmelt runoff event every spring.

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