EQUIVALENT PERMEABILITY OF A CONTINENTAL, ALPINE SNOWPACK

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ABSTRACT

Most snowpacks are not homogeneous, but consist of a number of snow and ice layers with varying permeabilities. In such cases, an 'equivalent' permeability can be used as an integrated or bulk property of the entire snowpack depth and is a useful parameter for gravity-flow and numerical models of meltwater flow through isothermal snowpacks. Calculated surface meltwater flux and lysimeter discharge were used to establish daily relationships between specific meltwater fluxes and wavespeeds through the snowpack at a continental, alpine site in 1996 and 1997. These relationships were used to determine a pore-size distribution index (ε) and to calculate equivalent permeability. Values for ε were found to vary considerably and could not be related to other measured snowpack properties. Values were in the lower range of those previously reported, often less than 3, and were highly dependent on the range of meltwater fluxes used in calculations. Equivalent permeability was also at the lower limit of values reported in the literature, and varied by 2 orders of magnitude from 1.08×10^{-11} to 2.79×10^{-8} m^2. In 1997 it declined considerably over 16 days.

INTRODUCTION

Movement of water through snowpacks is one of the least understood aspects of snow hydrology (Richter-Menge and Colbeck, 1991). It has an important influence on the timing and magnitude of snowmelt hydrographs (e.g. Caine, 1992), and on biogeochemical and geomorphological processes (e.g. Williams and Melack,1989; Caine, 1995). A considerable amount of research has been undertaken to investigate the underlying physics controlling water movement through snowpacks (e.g. Colbeck, 1973, 1972; Ambach et al., 1981; Jordan, 1983; McGurk and Kattelmann, 1988). This work has revealed a complex series of processes involving snowpack thinning, water flow through variably saturated porous medium, and refreezing of meltwaters within the snowpack, which causes changes in its hydraulic and thermal properties. These studies have also shown that snowpacks are highly heterogeneous due to microscale variations in density and ice grain structure, and macroscopic layering. However, there are still few experimental studies which quantify these processes under natural field conditions.

It is not clear how to determine an appropriate value of intrinsic (saturated) permeability (κ_s) for layered snowpacks. Point measurements of κ_s have little meaning if the permeability is used to predict the rate at which meltwater moves through the snowpack, encountering numerous layers with their associated individual permeabilities. Attempts have been made to model meltwater flux through a heterogeneous, layered snowpack by a variety of approaches, including; (i) modeling flow around individual ice layers (Colbeck, 1973); (ii) simplifying the effect of ice layers and preferential flowpaths by using multiple flowpath models (Marsh and Woo, 1985); and (iii) treating the snowpack as a homogeneous anisotropic media (Colbeck, 1975). Such models provide insight into flow processes within the snowpack, but the lack of data about snowpack horizons and layers makes them impractical in nearly all circumstances. The vast majority of studies, whether analytical or numerical, are forced to treat the snowpack as an equivalent, uniform medium in which permeability is considered as a bulk or integrated property of entire snowpack depth. McGurk and Kattelmann (1986) suggest that an equivalent permeability (κ_e) is very different from point values of intrinsic permeability, but is a very useful parameter for porous media gravity-flow (e.g. Colbeck, 1972, 1979; Dunne et al., 1976; Marsh and Woo, 1985) and numerical models (e.g. Jordan, 1983) of meltwater flow through isothermal snowpacks.

Unlike soils, laboratory measurements of snow permeability are almost impossible, because of the triple phase nature of snowpacks, and constant changes in density, grain size, type and distribution throughout the melt season due to metamorphism. Measuring κ_e in the field overcomes these problems, and can detect changes which occur during the melt season. The only method for determining κ_e uses measurements of surface meltwater flux (U) and the velocity at which this flux moves through the snowpack ($[dz/dt]_U$) (Marsh, 1991). When these data are plotted for periods of declining flux, the intercept of the line can be used to determine a value for κ_e over the entire

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snowpack depth, and the slope of the line can be used to determine the pore-size distribution index (ε) of the snow. ε is the exponent in the relationship between relative permeability (κ_w), κ_s and effective saturation (S*) for unsaturated snow such that as ε increases κ_w decreases. ε has been determined from drainage experiments, with reported values ranging between 1.5 and 4.8 (Marsh, 1991). κ_s or κ_e values for snow have been determined using a variety of methods, including the drainage method used here (i.e. Colbeck and Davidson, 1973; Colbeck and Anderson, 1982; Denoth et al. 1979a, b; Ambach et al., 1981) with reported values ranging between 6 x 10^{-11} and 3 x 10^{-8} m^2 (McGurk and Kattelmann, 1986).

Both ε and κ_e depend upon physical properties of the snowpack, and may change considerably during the melt season as snowpack metamorphism affects grain size, density and ice layer thickness. However, previous work has not been able to establish any relationships between ε and κ_e and snowpack properties or amounts of metamorphism (Marsh, 1991), although it has been suggested that κ_e is dependent upon the range of U used in calculations (McGurk and Kattelmann, 1986).

Surface meltwater flux, lysimeter discharge and snowpack properties are used in this paper to investigate (1) whether ε and κ_e can be related to more easily measured snowpack properties and state of metamorphism, and (2) whether they vary consistently in a way which can be used to parameterize gravity-flow and numerical models during the melt season.

METHODS

Site Description

Data were collected in the melt seasons of 1996 and 1997 from a lysimeter array located on Niwot Ridge, in the Colorado Front Range of the Rocky Mountains, about 5km east of the Continental Divide (40° 03"N, 105° 35"W). The saddle site, within the Niwot Ridge Long Term Ecological Research (NWTLTER) is an alpine tundra environment at an elevation of 3515m and is located in a relatively flat, broad saddle. Since 1951, mean annual temperature has been –3.8°C and annual precipitation is 1006mm (Williams et al., 1996). The depth of snow accumulation varies greatly along the ridge, due to interactions between local topography and high wind velocities, with wind scoured bare ground found adjacent to areas of accumulation with depths of over 8m.

Snowpack properties

Snow depth was monitored constantly beneath the instrument tower using a Campbell Scientific UDG01 Ultrasonic Depth Gauge. This sensor measures the distance from the sensor to the surface; it is currently mounted at 6.0m above the ground, so during peak snowpack conditions it was 2-3 m above the snow surface. A systematic weekly program of snowpit excavation approximately 100 m from the lysimeter array recorded vertical profiles of temperature and density, measured with a 0.001m^3 steel cutter at 0.1m intervals following the protocol of Elder et al. (1991). The depth and thickness of any ice layers was recorded, and the grain size and type for individual snow horizons were noted.

Snowmelt season

Maximum snow depths of 275 cm were recorded on JD 116 (25 April), 1996 and 225 cm on JD 122 (2 May), 1997. In both years snow melt rates were relatively low for the first 30 days after peak accumulation. Snowpack depths were reduced by approximately 70 cm over this period, with additional accumulation recorded around JD 150 in both 1996 and 1997. After this period, snow melt rates more than doubled and remained high until complete ablation had occurred by JD 180 (28 June), 1996 and by JD 178 (26 June), 1997. Maximum melt rates were similar in both years at 5.5 mm hr^{-1}, although there was considerably day to day variation in the patterns of diurnal melt.

Energy balance

Surface meltwater flux (U) was calculated from direct measurements of the snow-atmosphere energy exchange and internal energy balance of the snowpack using the protocol in Cline (1997) and data presented in Williams et al. (1999, in press). The accuracy of the instruments and their proximity to the lysimeters meant that surface meltwater flux could be modeled with good precision over a wide range of fluxes and short time intervals. This represented a considerable improvement over previous studies.

Lysimeter discharge

Snowmelt water draining from the base of the snowpack was measured using lysimeters located on the ground surface about 3m from the meteorological instrument tower. Two 1 m^2 lysimeters drain into a subnivean laboratory where discharge was measured using Campbell Scientific TE525 Tipping Bucket Rain Gages with 10 minute totals recorded on a Campbell CR21x data logger which were aggregated to hourly totals for this analysis. (For a full description of the NWTLTER saddle lysimeter array see Rikkers et al., 1996).

Wavespeeds ($[dz/dt]_U$)

For downward, non-saturated flux of water ignoring water pressure gradients Colbeck (1974) found any surface melt flux (U) propagates through the snowpack at speed $[dz/dt]_U$:

$$\left[\frac{dz}{dt}\right]_U = \frac{\varepsilon\,\alpha^{1/\varepsilon}\,\kappa_s^{1/\varepsilon}\,U^{(\varepsilon-1)/\varepsilon}}{\phi^*} \qquad (1)$$

Where: ε = pore size distribution index; κ_s = saturated, or intrinsic, permeability; ϕ^* = effective porosity, α = flow factor, 5.47 x 10^6m-s^{-1}.

The speed ($[dz/dt]_U$) at which a certain surface meltwater flux (U) moves down through the snowpack was determined in a non-destructive way by timing how long (dt) it took U to travel the entire depth of the snowpack (dz). U was calculated for a given time step from the meteorological data, and then lysimeter discharge (Q) was used to detect when this flux reached the base of the snowpack. To identify when a particular U reached the snowpack base, calculated U and Q were converted into cumulative frequency curves, which effectively forced the volumes to match each other, allowing comparison. The time lag between cumulative U at a particular time step, and the same amount of cumulative Q was calculated. In every case, $[dz/dt]_U$ was calculated from the recession limbs of the diurnal lysimeter hydrographs, when Q was decreasing, with drainage of meltwater from the snowpack due to gravity in an unsaturated flow regime.

Pore-size distribution index (ε)

The theoretical pore-size distribution index (ε) for the snowpack was calculated for each day on which $[dz/dt]_U$ were calculated. The slope (b) of log-log plots of $[dz/dt]_U$ against U gives the flux exponent in equation 1, (ε-1/ε). If ε = 3, then b should be 0.67. Figure 1 for JD 168, 1997 shows 8 values of U, b = 0.525, ε = 2.11, R^2 = 0.93.

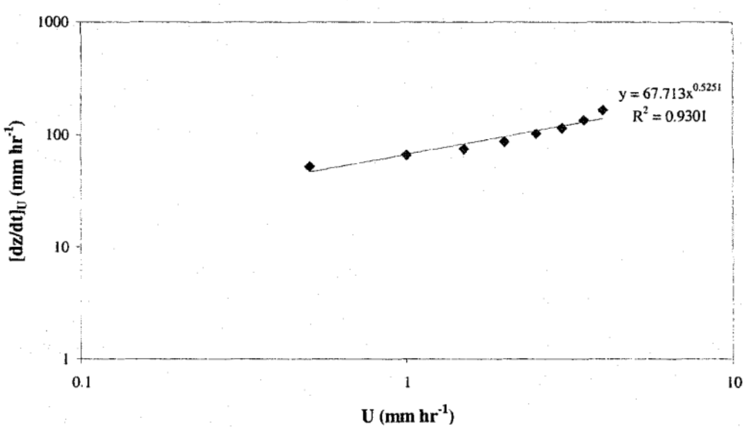

Figure 1. Wavespeed ($[dz/dt]_U$) regressed on surface meltwater flux (U) ranging between 0.5 and 4 mm hr^{-1} for data from lysimeter 2, JD 168 1997.

Equivalent permeability (κ_e)

For each day analyzed, mean U for all values used in calculating the $[dz/dt]_U$ against U regression was substituted into the regression equation to give $[dz/dt]_{mean\ U}$. These values of mean U and $[dz/dt]_{mean\ U}$ were then substituted into equation 1. Given a value for ε, and that α is a constant, rearranging equation 1 gives $\kappa^{1/\varepsilon}\phi^{*-1}$. This single value characterizes an important property of the snowpack, which Colbeck (1978) suggested should have typical values assigned, and then be used as a parameter in operational forecasting, changing as the snowpack matures. Having calculated effective porosity (ϕ^*) for a given day using density and irreducible water content values (McGurk and Kattelmann, 1986), κ_e can be determined from $\kappa^{1/\varepsilon}\phi^{*-1}$.

RESULTS AND DISCUSSION

Data analyzed here were collected in two melt seasons: JD 155 (3 June) to 172 (20 June), 1996; and JD 150 (30 May) to 175 (24 June), 1997.

Wavespeeds ($[dz/dt]_U$)

Table 1 summarizes log-log plots of $[dz/dt]_U$ against U for 9 days during 1996 and 10 days during 1997. Days were selected for analysis when daily cumulative melt most closely matched daily cumulative Q. This was generally later in the melt seasons when there was large quantities of melt, and Q rates rose well above the minimum value used in analysis, 0.5 mm hr^{-1}. a and b are the exponents in the regression equation, $y = a(x)^b$, for $[dz/dt]_U$ regressed against U and $[dz/dt]_{2\ mm\ hr\text{-}1}$ values represent normalized wavespeeds calculated using U = 2 mm hr^{-1} allowing comparison over the melt season.

Table 1. Summary of $[dz/dt]_U$ regressed on U showing day, lysimeter, number of U used to calculated $[dz/dt]_U$ for that day, daily mean U, R^2 for the relationship between U and $[dz/dt]_U$, exponents of the equation of this relationship, and calculated $[dz/dt]_{2\ mm\ hr\text{-}1}$.

Year	Day	Lysimeter	n	Mean U (mm hr^{-1})	R^2	a	b	$[dz/dt]_{2\ mm\ hr\text{-}1}$ (mm hr^{-1})
1996	156	1	6	1.75	0.890	115.410	0.154	128.402
1996	157	1	8	2.75	0.556	112.370	0.155	125.098
1996	158	1	8	2.50	0.830	118.250	0.376	153.458
1996	159	1	8	2.50	0.799	122.260	0.422	163.825
1996	160	1	8	2.50	0.904	134.260	0.240	158.582
1996	163	1	8	2.50	0.839	115.950	0.184	131.686
1996	169	1	8	2.75	0.700	74.328	0.681	119.191
1996	170	1	8	2.75	0.847	91.165	0.816	160.475
1996	171	1	8	2.75	0.591	69.709	0.716	114.537
1997	155	1	8	2.25	0.698	87.311	0.477	121.498
1997	156	1	8	2.25	0.882	108.290	0.309	134.136
1997	158	2	8	1.13	0.883	94.938	0.486	133.003
1997	160	1	5	0.75	0.525	82.045	0.135	90.081
1997	162	1	8	2.25	0.521	65.992	0.540	95.977
1997	165	2	8	2.25	0.654	62.240	0.821	109.947
1997	168	1	8	2.25	0.916	71.858	0.520	103.063
1997	168	2	8	2.25	0.930	67.713	0.525	97.441
1997	171	1	8	2.25	0.717	38.943	0.699	63.224
1997	171	2	8	2.25	0.699	38.611	0.674	61.612

In general, R^2 were high, indicating in all cases that variance in U controls over 50% of variance in $[dz/dt]_U$. In all cases b was positive, indicating that increasing U resulted in increasing $[dz/dt]_U$, but there was considerable variation in the actual value, ranging from 0.13 for lysimeter 1, JD 160 1997 to 0.82 for lysimeter 2, JD 165 1997, with a mean of 0.47, and a standard deviation (n = 19) of 0.228. This suggests that the relationship between U and $[dz/dt]_U$ varies during the melt season. $[dz/dt]_U$ was calculated for U = 2 mm hr^{-1} using the equation for the regression line. A linear regression equation of the 1997 data indicated a clear decrease in wavespeeds as the melt season progresses ($R^2 = 0.67$) (Figure 2). In 1996 there is no equivalent seasonal trend ($R^2 = 0.06$)

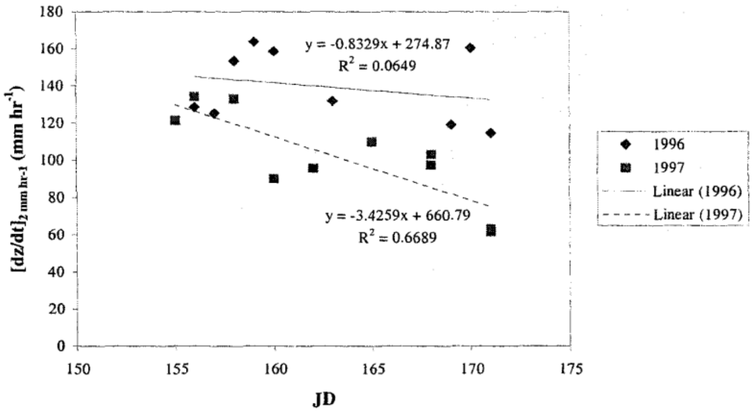

Figure 2. Wavespeed calculated for surface meltwater flux of 2 mm hr^{-1} ($[dz/dt]_{2\ mm\ hr-1}$) regressed on Julian day.

Although R^2 for many daily relationships between $[dz/dt]_U$ and U were high, in most cases all data did not lie on the best fit line. $[dz/dt]_U$ for higher U would fit better if the gradient of the line increased. This suggests that the relationship between $[dz/dt]_U$ and U is dependent on the range of U selected to establish the relationship. To test this the data were split into 2 approximately equal sized groups for each year (Table 2).

Table 2. Summary of $[dz/dt]_U$ regressed on U for a variety of U ranges, showing year, range of U, the number of U in that range, mean U in that range, R^2 for the relationship between U and $[dz/dt]_U$, and the exponents of the equation of this relationship.

Year	U range (mm hr^{-1})	n	Mean U (mm hr^{-1})	R^2	a	b
1996	<2.5	45	1.5	0.1888	104.79	0.1872
1996	>2.5	37	3.86	0.3927	30.772	1.3844
1996	all	82	2.57	0.4880	103.44	0.4876
1997	< 2	36	0.97	0.0665	68.436	0.1809
1997	>2	41	2.98	0.4504	33.743	1.1938
1997	all	77	2.04	0.4310	72.727	0.4541

These values can be compared to regressions between $[dz/dt]_U$ and U values > 1 mm hr-1 in both years, when in 1996, $y = 70.261\ x^{0.7708}$ (n = 64, $R^2 = 0.49$), and in 1997, $y = 50.688\ x^{0.8386}$ (n = 53, $R^2 = 0.41$). These indicated that when smaller U values were excluded, the gradient of the regression lines increased significantly, suggesting that the nature, and possibly the strength, of the $[dz/dt]_U$ – U relationship was dependent upon the range of U chosen.

$[dz/dt]_U$ were similar in both 1996 and 1997, although they declined with time in 1997 and remained constant over a relatively short study period during 1996. The values calculated for U = 2 mm hr^{-1} were at the lower limit of those reported for late melt season in the literature, and much lower than those reported for earlier in the melt season. Jordan (1983) reported $[dz/dt]_U$ ranged between 108 and 360 mm hr^{-1} for U of 1 to 3 mm hr^{-1}, and similarly McGurk and Kattelmann (1986) found U of 2 mm hr^{-1} to have a wavespeed of 180 mm hr^{-1} in May at a Sierra Nevada site but was twice that in March. The approximate 50% reduction in $[dz/dt]_U$ reported here for 1997 is of a similar magnitude but occurred in only 16 days.

Pore-size distribution index (ε)

ε was calculated from the slope (b) (Table 2) of the regression line for each log-log plot of $[dz/dt]_U$ against U. Table 3 shows calculated ε, and $\kappa^{1/\varepsilon}\,\phi^{*-1}$ and κ_e calculated with ε held constant as 3.

Table 3. Year, day, lysimeter, number of U used to calculated $[dz/dt]_U$ on that day, mean U, R^2 for relationship between U and $[dz/dt]_U$, $[dz/dt]_U$ calculated using mean U and calculated ε, $\kappa^{1/\varepsilon}\,\phi^{*-1}$, ϕ^* and κ_e.

Year	Day	Lysimeter	n	Mean U (mm hr^{-1})	R^2	ε	$[dz/dt]_{meanU}$ (mm hr^{-1})	$\kappa^{1/\varepsilon}\,\phi^{*-1}$ (m$^{2/3}$)	ϕ^*	κ_e (m^2)
1996	156	1	6	1.75	0.890	1.182	125.790	1.07E-03	0.462	1.21E-10
1996	157	1	8	2.75	0.556	1.183	131.419	8.26E-04	0.462	5.58E-11
1996	158	1	8	2.5	0.830	1.603	166.889	1.12E-03	0.462	1.38E-10
1996	159	1	8	2.5	0.799	1.731	180.009	1.21E-03	0.462	1.73E-10
1996	160	1	8	2.5	0.904	1.316	167.314	1.12E-03	0.462	1.39E-10
1996	163	1	8	2.5	0.839	1.225	137.193	9.19E-04	0.496	9.46E-11
1996	169	1	8	2.75	0.700	3.138	148.071	9.31E-04	0.496	9.82E-11
1996	170	1	8	2.75	0.847	5.429	208.083	1.31E-03	0.449	2.02E-10
1996	171	1	8	2.75	0.591	3.526	143.889	9.05E-04	0.449	6.69E-11
1996	All		82	2.57	0.488	1.757	155.327	1.02E-03	0.467	1.08E-10
1996	<2.5		45	1.5	0.189	1.230	121.969	1.15E-03	0.467	1.54E-10
1996	>2.5		37	3.86	0.393	-2.60	1295.002	6.50E-03	0.467	2.79E-08
1996	>1		64	3.08	0.509	4.363	397.983	2.32E-03	0.467	1.27E-09
1997	155	1	8	2.25	0.698	1.911	128.515	9.24E-04	0.472	8.28E-11
1997	156	1	8	2.25	0.882	1.447	139.105	1.00E-03	0.472	1.05E-10
1997	158	2	8	1.125	0.883	1.947	100.536	1.15E-03	0.512	2.03E-10
1997	160	1	5	0.75	0.525	1.156	78.924	1.18E-03	0.512	2.21E-10
1997	162	1	8	2.25	0.521	2.176	102.285	7.35E-04	0.512	5.35E-11
1997	165	2	8	2.25	0.654	5.583	121.109	8.71E-04	0.489	7.71E-11
1997	168	1	8	2.25	0.916	2.085	109.576	7.88E-04	0.489	5.71E-11
1997	168	2	8	2.25	0.930	2.106	103.658	7.45E-04	0.489	4.84E-11
1997	171	1	8	2.25	0.717	3.323	68.650	4.94E-04	0.461	1.18E-11
1997	171	2	8	2.25	0.699	3.069	66.704	4.80E-04	0.461	1.08E-11
1997	All		77	2.04	0.431	1.832	100.551	7.72E-04	0.487	5.31E-11
1997	< 2		36	0.97	0.067	1.221	68.060	8.57E-04	0.487	7.28E-11
1997	>2		41	2.98	0.450	-5.16	124.252	7.41E-04	0.487	4.69E-11
1997	>1		53	2.64	0.400	6.196	114.410	7.39E-04	0.487	4.67E-11

In every case, $[dz/dt]_U$ was calculated from the recession limbs of the diurnal lysimeter hydrographs. However, b from the regression line for the $[dz/dt]_U$ – U relationship calculated for a range of U >2.5 in 1996 and >2 in 1997 were larger than 1.0. As values greater than 1.0 indicate increasing meltwater flux, and the formation of a 'shock wave', which was not the case at these times on any day, these high values of b were assumed to be an artifact of the way the relationship was established from a range of U chosen arbitrarily. The resulting ε were considered to be erroneous and were excluded from statistical analysis and further discussion.

19 ε values were calculated from exponents for daily $[dz/dt]_U - U$ regression lines. These ranged between 1.16 and 5.58, with a mean value of 2.38 and a standard deviation (n = 19) of 1.34. When the additional 6 values 'All', '<2.5', '<2' and '>1' were included, the maximum ε increased to 6.20, whilst the mean increased slightly to 2.47, and the standard deviation (n = 25) increased to 1.50.

In order to investigate whether ε showed a seasonal trend over the melt season, perhaps reflecting metamorphism and density changes within the snowpack, ε was regressed on date for each of the two years. Figure 3 shows that ε in 1996 increase consistently with time at the rate of 0.212 per day ($R^2 = 0.73$). In 1997, R^2 was reduced to only 0.22 as the relationship was especially affected by JD 165 value. When this was excluded ε exhibited a smaller increase of 0.083 per day ($R^2 = 0.59$). There was considerable variation in ε values over these relatively short periods, and even successive days had very different values.

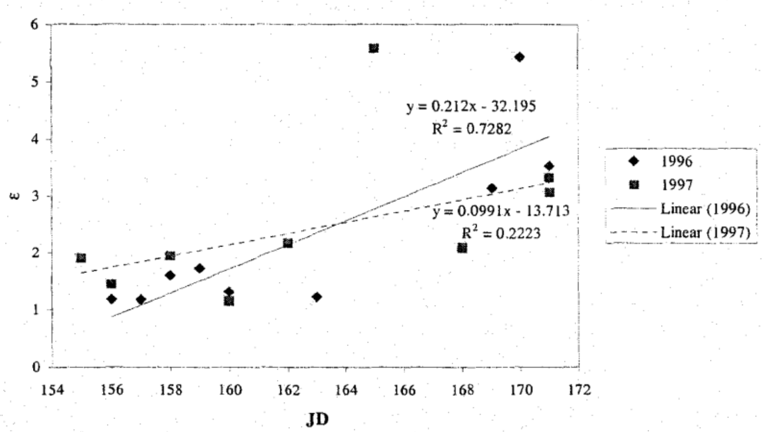

Figure 3. ε regressed on Julian day (JD) for 1996 and 1997.

As ε represents a physical property of the snowpack, an attempt was made to relate ε to those snowpack properties which were measured in the field, density, grain size and ice layer thickness. No trends were apparent for density (1996, $R^2 = 0.10$; 1997, $R^2 = 0.07$). For both years ε increased as grain size increased (1996, $R^2 = 0.15$; 1997, $R^2 = 0.27$), and increased as the total thickness of the ice layers in the snowpack decreased (1996, $R^2 = 0.30$; 1997, $R^2 = 0.14$), but as there was considerable scatter in the results there were no significant relationships. Of these measured snowpack properties, only mean grain size should be related directly to ε as it is well established that ϕ^* (established from density) reveals nothing about ε (Hillel, 1971). Which, if any, of these measured snowpack properties controls ε remains unknown as they all vary with time.

As ε was determined directly from b which seemed to be dependent on the range of U used when calculating the regression, ε must also be dependent on U. Calculated ε was regressed against mean U for individual days in 1996 and 1997. However, many of the cases had the same mean U limiting the explanatory value of the plot, but the linear regression equations for 1996 and 1997 each had a R^2 of 0.31, and a trend of increasing ε with mean U seemed apparent.

ε calculated in this study was generally lower than the commonly used value of 3, with a mean below 2.5. However, it was clear that when higher ranges of U were used the slope of the regression line increased and ε was calculated to be considerably higher than 3. As ε is a physical property of the snowpack it may be expected to change over the melt season but should not vary in response to the range of U used to calculate it. That it did suggested that equation (1) does not provide an adequate explanation of the complex processes it attempts to model at all ranges of U, and this causes inaccuracies when calculating ε using this method. 2.5 is the minimum ε report for sand by Mualem (1978), and lies towards the low end of values reported for snow. Although ε varied considerably between days, there was an increasing trend over the melt season (Figure 3). However, ε could not be related definitely to other measured snowpack properties which were found to vary over the melt season, and it is possible that the apparent trend in ε resulted from small variations in the range of U used on different days (Table 3), which changed most during 1996.

Equivalent permeability (κ_e)

In order to allow comparison with values of κ_e reported in the literature κ_e was determined both for calculated ε, and when ε was held constant as 3 (Table 3). When ε was held constant, κ_e ranged between 1.08 x 10^{-11} and 2.79 x 10^{-8} m^2, with a mean = 1.17 x 10^{-9} m^2, and standard deviation 5.39 x 10^{-9} m^2 (n = 27), although these statistics were heavily influenced by two relatively high values for '>2.5' and '>1' in 1996. If these were excluded mean κ_e fell to only 9.76 x 10^{-11} m^2. When κ_e was calculated using our estimated values of ε it varied from day to day over several orders of magnitude. Holding ε constant as 3 greatly reduced the variability, and gave κ_e at the low end of reported values, similar to those for fine grained snow.

Figure 4 is a plot of κ_e against JD for values calculated in 1996 and 1997. In 1996 there was no visible trend, but in 1997 κ_e declined over time (R^2 = 0.44). Similarly, κ_e in 1996 showed no significant relationships to snowpack density, grain size and total ice layer thickness. In 1997 κ_e was inversely related to density (R^2 = 0.47) and mean grain size (R^2 = 0.20), and directly to total ice layer thickness (R^2 = 0.22).

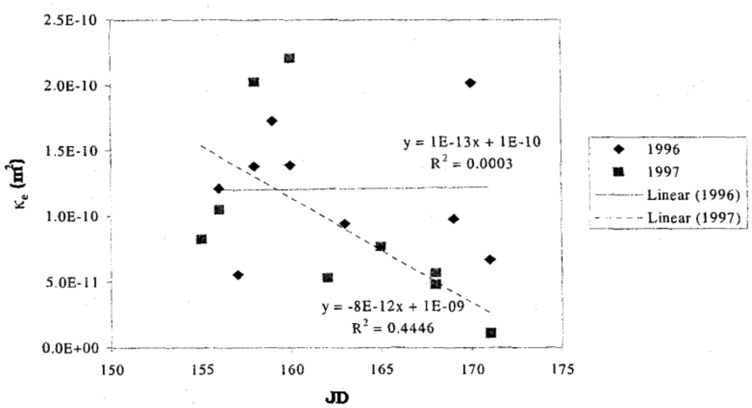

Figure 4. Equivalent permeability (κ_e) regressed on Julian day (JD) for 1996 and 1997.

Figure 5 plots κ_e and κ_s against JD for 1996 and 1997. It shows an order of magnitude difference between κ_e and κ_s, (κ_s calculated from density and grain size as Shimizu (1970) and κ_e from wavespeed analysis). In 1996 neither κ_e nor κ_s showed any significant trends over the season, but in 1997 κ_e decreased with time (R^2 = 0.44) and κ_s increased (R^2 = 0.69). This emphasizes that κ_e and κ_s are measures of very different snowpack properties.

As discussed, $[dz/dt]_U$ and ε seemed to be dependent upon the range of U used to calculate them. It should follow that κ_e was similarly dependent. Again, due to the limited number of mean U, trends in the data were not very distinctive. In 1996 κ_e seemed to increase with increasing mean U, but the opposite was true in 1997.

It is possible that the change in κ_e was a response to changing ice layer thickness and number. Early in the melt season relatively impermeable ice layers caused lateral flow and concentration of meltwater into preferential pathways where higher levels of saturation increased hydraulic conductivity and effective permeability. As the ice layers began to breakdown these preferential pathways degenerated, resulting in more uniform, slower meltwater movement through the snowpack and reduced levels of saturation.

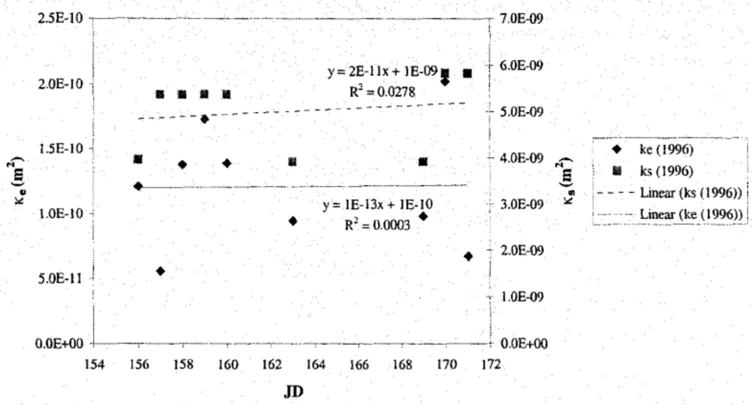

Figure 5 (a). Equivalent permeability (κ_e) and intrinsic permeability (κ_s) regressed on Julian day (JD) for 1996.

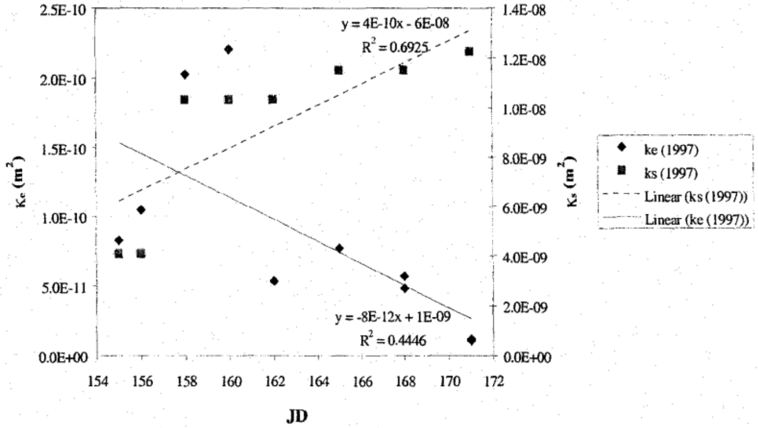

Figure 5 (b). Equivalent permeability (κ_e) and intrinsic permeability (k_s) regressed on Julian day (JD) for 1997.

CONCLUSIONS

Using calculated meltwater flux and lysimeter discharge for 2 melt seasons at a continental, alpine site, meltwater wavespeeds were found to be approximately 120 mm hr^{-1} for a surface meltwater flux of 2 mm hr^{-1}. This is similar to lower values reported in the literature. Wavespeeds were found to decrease by 50% over only 16 days in 1997 while snowpack depth decreased from 130 to 40 cm over the same period. The pore-size distribution index (ε) was calculated using these wavespeed data and was found to be generally lower than the commonly used value of 3, with a mean value below 2.5. However, it was very sensitive to the surface meltwater flux range used in calculating wavespeeds. This suggests that the processes controlling meltwater movement through a snowpack are more complex than simple gravity-flow in an unsaturated porous medium. ε could not be related definitely to any measured snowpack properties. κ_e was also found to be at the lower limit of reported values, with a mean value of 1.17×10^{-9} m^2, and many individual estimates over an order of magnitude smaller. Equivalent permeability declined considerably during 1997 in contrast to increasing calculated intrinsic permeability. This possibly reflects the degeneration of preferential flowpaths and areas of concentrated meltwater movement earlier in the melt season.

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