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Isotopic Composition of Vent Discharge from the Matanuska Glacier, Alaska: Implications for the Origin of Basal Ice

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Daniel D. Titus

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ISOTOPIC COMPOSITION OF VENT DISCHARGE FROM THE MATANUSKA GLACIER, ALASKA: IMPLICATIONS FOR THE ORIGIN OF BASAL ICE.

By

Daniel D. Titus

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Submitted to
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ABSTRACT

ISOTOPIC COMPOSITION OF VENT DISCHARGE FROM THE MATANUSKA GLACIER, ALASKA: IMPLICATIONS FOR THE ORIGIN OF BASAL ICE.

By

Daniel D. Titus

Basal ice at the Matanuska Glacier, Alaska is characterized by elevated ^{3}H concentrations and less negative $\delta^{18}O$ and δD values relative to englacial ice. Recently it has been suggested that basal ice is partially composed of modern meteoric water and forms as subglacial discharge moves out of an overdeepening resulting in supercooling of the discharge and nucleation of ice crystals.

Values of $\delta^{18}O$, δD and ^{3}H were measured in subglacial discharge collected during the summer of 1995 from vents along the ice margin at the Matanuska Glacier. Application of a simple open system freezing fractionation model indicates that ^{18}O, D and ^{3}H in discharge are within the requisite ranges to form basal ice, and that there is a genetic relationship between the two. Additionally, the temporal variability of $\delta^{18}O$ in the basal ice can be attributed to annual deviations in relative amounts of precipitation and meltwater present in subglacial discharge.

To my loving wife and confidant Suzanne, you are the joy in each moment that is forever my source of motivation. Your strength and devotion are unrivaled, and compel me every day of my life. I Love you.

ACKNOWLEDGMENTS

The work presented in this manuscript, and indeed the completion of my graduate degree, would not have been possible without the patient tutelage of my advisor Dr. Grahame J. Larson. His confidence in me and my ability has fostered intellectual and professional growth beyond my greatest expectations. Dr. Larson has proven time and time again to be a consummate mentor and professor, not to mention a great friend. Thank you for everything Grahame.

I would also like to thank my parents Roger and Faye Titus. Their contribution is not obvious in the text contained here in, but it is pervasive in all that I do and have become. They taught me how to strive to live up to my own expectations through perseverance and hard work. To this day, I believe that is the greatest lesson I have ever learned. Thanks Mom and Dad.

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INTRODUCTION

Traditionally, formation of the debris rich basal ice zone observed in many temperate and subpolar glaciers has been attributed to the regelation process (Weertman, 1961, 1964; Boulton, 1972; Iverson, 1993). Regelation occurs as the result of phase changes caused by either variations in temperature at the glacier sole or pressure fluctuations in a glacier that is near its pressure-melting point. These phase changes may occur in response to: isolated cold patches or permafrost at the glacier sole (Robin, 1976), as cold winter air penetrates to the sole (Weertman, 1961; Clarke et al., 1984), or through melting on the upstream side and subsequent refreezing on the downstream side of an obstacle encountered by moving ice at the base of the glacier (Kamb and LaChapelle, 1964; Iverson and Semens, 1995). Irrespective of the particular cause of the phase change, clean englacial ice is melted and then refrozen thereby incorporating basal debris. Physical constraints on the regelation process dictate, however, that this mechanism cannot incorporate a net thickness of basal ice beyond $1cmy^{-1}$ (Alley et al., 1996) Thus, regelation alone cannot explain the >1m exposures of basal ice observed at the Matanuska Glacier, Alaska (Lawson et al. 1996; Stasser et al., 1996).

Alternatively it has been shown on theoretical grounds when water moving through an open linked-cavity drainage system flows out of an overdeepening, at or near the terminus of a glacier, a net accretion of ice through freeze-on will occur at the glacier bed if the aspect ratio of the bed to surface gradient is between -1.2 and -1.7 (Figure 1) (Alley et al., in press). This freeze-on process occurs because subglacial discharge flowing up gradient out of the overdeepening becomes supercooled due to a rapidly increasing pressure melting point and insufficient influx of heat to compensate for the pressure change (Strasser et al., 1996; Lawson, in review; Alley et al., in press). Supercooling of the discharge results in nucleation of ice crystals in the water and thereby the release of latent heat, thus allowing the water to maintain thermal equilibrium. According to this hypothesis, ice crystals formed in the supercooled discharge are subsequently affixed to drainage cavity surfaces resulting in a net accretion of basal ice (Strasser et al., 1996; Lawson et al., in review; Alley, in press).

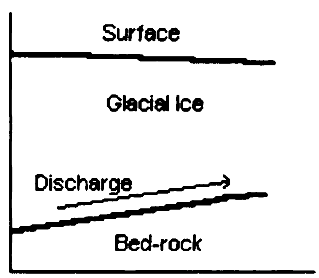

Figure 1. Aspect Ratio Of Surface To Bed Required For Freeze-on

Strasser et al.(1996) and Lawson et al. (in review) recently observed that $\delta^{18}O$ values in frazzil ice and basal ice at the Matanuska Glacier are virtually identical and enriched greater than 3‰ in ^{18}O relative to englacial ice. Frazzil ice forming around subgalcial discharge vents at the Matanuska Glacier is attributed to supercooling of discharge, thus Strasser et al.(1996) and Lawson et al. (in review) conclude that frazzil ice and basal ice are both formed by the freeze-on process. Furthermore, because the maximum enrichment of ^{18}O in response to fractionation during freezing is only 3‰ (O'Niel, 1968; Lehmann and Siegenthaler, 1991), they additionally conclude that the source water for frazzil ice and basal ice is not comprised solely of englacial melt water, but of a mixture of meltwater and isotopically heavier meteoric water (Strasser et al., 1996). This conclusion was supported further by the observation that much of the basal ice contains appreciable amounts of bomb ^{3}H, 5->95 TU, which is possible only if the source water for the basal ice is partially derived from recent meteoric water (Strasser et al., 1996; Strasser, 1995).

The hypothesized basal ice freeze-on theory requires, therefore, that the isotopic composition of vent discharge at the terminus of the Matanuska Glacier (corrected for fractionation due to freezing) reflect the composition and variation of basal ice with respect to it's ^{18}O, D, and ^{3}H content. In this paper, we present results of ^{18}O, D and ^{3}H measurements in vent discharge collected along the ice margin at the Matanuska Glacier, and attempt to show that there is a genetic relationship between vent discharge and basal ice.

DEVELOPMENT

Based on the physical parameters of a simple closed reservoir system, Jouzel and Souchez (1982) derived equations to describe the fractionation path of ^{18}O and D during freezing of subglacial water. In this closed system model the physical parameters of the reservoir are defined as follows: Input to the reservoir (I) is equal to zero; Output (O) from the reservoir is equal to zero; The freezing rate of the reservoir (F) is less than the isotopic diffusion coefficient (e.g. 10^{-5}cms^{-1}) eliminating the possibility of isotopic gradients in the water; Finally, δR_I is the initial $\delta^{18}O$ and δD values of the subglacial reservoir. The $\delta^{18}O$ and δD values of basal ice formed in this conceptual closed reservoir system are given by the equations (Jouzel and Souchez, 1982):

$$\delta_s{}^{18}O = 10(1000+\delta_i{}^{18}O)((1.1-K)^{\alpha}-(1-K)^{\alpha})-1000 \qquad \textbf{(1)}$$

$$\delta_s D = 10(1000+\delta_i D)((1.1-K)^{\beta}-(1-K)^{\beta})-1000 \qquad \textbf{(2)}$$

where $\delta_s{}^{18}O$ and $\delta_s D$ are isotope values of basal ice formed at each progressive fraction of the initial reservoir frozen (e.g. 0.1-1) denoted by K, and where α and β are equilibrium fractionation coefficients for $^{18}O/^{16}O$ and D/H, 1.00291 and 1.0212 respectively

(Lehmann and Siegenthaler, 1991).

On a diagram where $\delta_s{}^{18}O$ and δ_sD values from equations 1 and 2 are plotted for corresponding values of K, and as the abscissa and ordinate respectively, Jouzel and Souchez (1982) showed that $\delta^{18}O$ and δD values evolve co-linearly and fall on a freezing line with a slope value less than that of the Global Meteoric Water Line (e.g. 8). Therefore, Jouzel and Shouchez (1982) conclude that when basal ice sampled from a natural closed subglacial system falls on a freezing line, the initial δR_I composition of the reservoir being frozen is defined by the intersection of the freezing line and the Universal Precipitation Line. Thus, the relationship between δR_I and the $\delta^{18}O$ and δD values of basal ice can be determined by the slope of the line basal ice samples fall on irrespective of the K value any individual ice sample represents (Jouzel and Souchez, 1982). This closed reservoir system freezing slope is given by the equation (Jouzel and Souchez, 1982):

$$S = ((\alpha-1)/(\beta-1))((1000+\delta_i D)/(1000+\delta_i{}^{18}O)) \qquad \textbf{(3)}$$

where $\delta_i D$ and $\delta_i{}^{18}O$ are initial reservoir values and α and β are maximum equilibrium fractionation coefficients for $^{18}O/^{16}O$ and $^{2}H/^{1}H$ respectively (e.g. α=1.00291 and β=1.0212). An example of a closed reservoir system freezing line for a hypothetical δR_I composition, within the range of $\delta^{18}O$ and δD values measured at the Matanuska Glacier, is given in Figure 2.

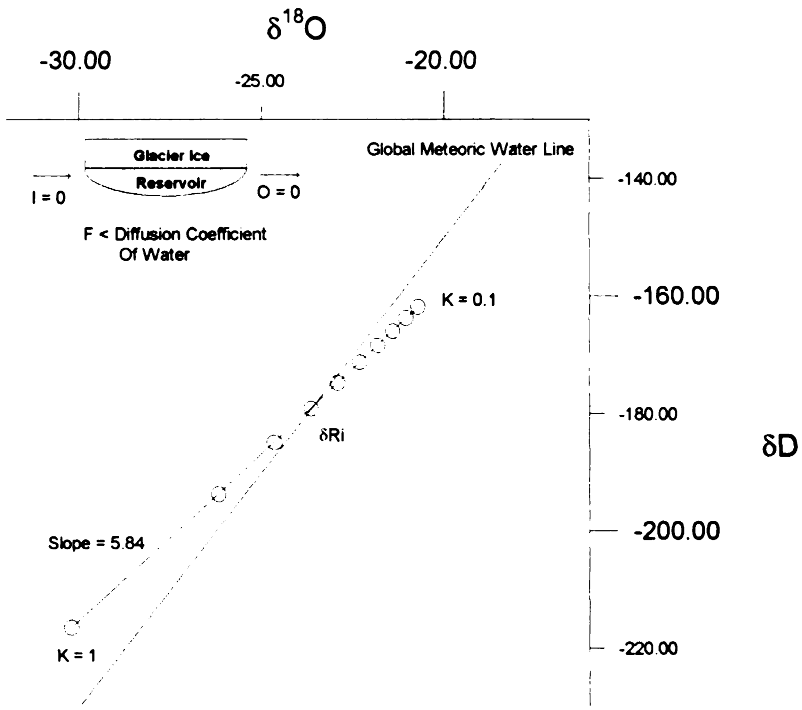

Figure 2. Closed Meltwater Reservoir

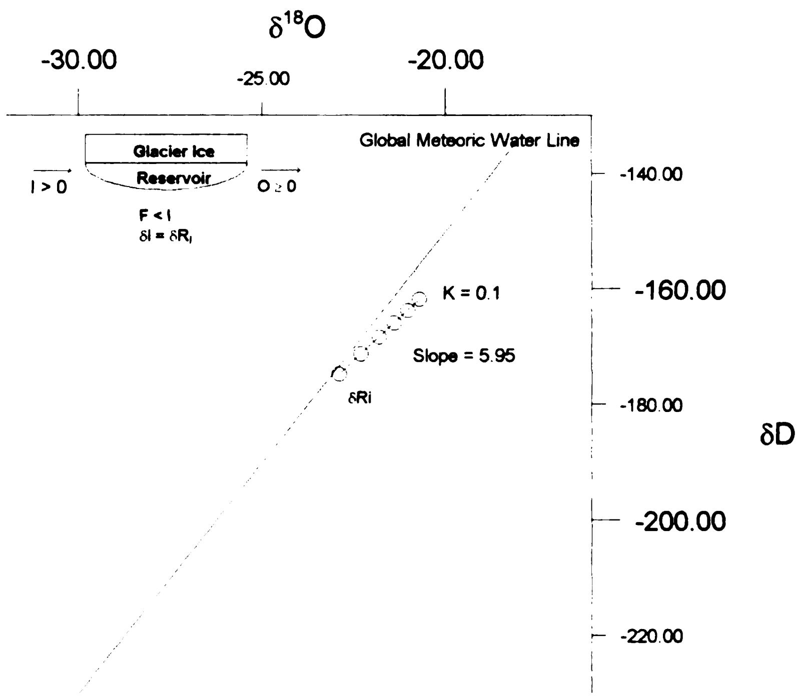

Figure 3. Open Meltwater Reservoir

Expanding on the conceptual closed reservoir system freezing model, Souchez and Jouzel (1984) developed a slope equation for interpretation of the relationship between δR_I composition and $\delta^{18}O$ and δD values of basal ice formed in an open meltwater subglacial reservoir. In this open meltwater system model the values of I, O, and F can vary according to the physical characteristics of the system being described, plus the isotopic value of the input (δI) is incorporated to account for changes in the ^{18}O and D content of I during freezing of the reservoir. Souchez and Jouzel (1984) showed when δI was not significantly different from δR_I, which is reasonable if δR_I and δI are composed primarily of englacial ice melt, that irrespective of the values of I, O and F the freezing slope in an open meltwater subglacial reservoir is given by the equation:

$$S = (\alpha/\beta)((\alpha-1)/(\beta-1))((1000+\delta_i D)/(1000+\delta_i{}^{18}O)) \qquad \textbf{(4)}$$

again where $\delta_i D$ and $\delta_i{}^{18}O$ are initial reservoir isotope values and α and β are maximum equilibrium fractionation coefficients for $^{18}O/^{16}O$ and $^2H/^1H$ receptively (e.g. α=1.00291 and β=1.0212). This open meltwater reservoir system slope equation yields a value that is virtually identical to that of the closed reservoir system freezing slope of Jouzel and Souchez (1982). Thus, for these open and closed system models the only significant difference in the freezing process is that ice produced in an open meltwater system is always enriched in ^{18}O and D relative to

δR_I due to continuous reconstitution of the reservoir by I, where δI and δR_I are equal and I is not less than F (Souchez and Jouzel, 1984). Alternatively, in an open meltwater system ice depleted in heavy isotopes relative to δR_I may be produced if the value of F is greater than I. This situation, however would be limited to systems that produce little meltwater such as high elevation cirque glaciers. An open meltwater reservoir system freezing line, where I is not less than F, is given in figure 3. The $\delta^{18}O$ and δD values used in this simulation are the same as those used previously in the closed reservoir system freezing line (Figure 2).

Hubbard and Sharp (1995) identified that single co-isotopic freezing slopes (Jouzel and Souchez, 1982; Souchez and Jouzel, 1984) were not adequate for interpretations of basal ice $\delta^{18}O$ and δD values if that ice was formed during discrete freezing episodes from reservoirs with different δR_I compositions. Furthermore, Hubbard and Sharp (1995) suggested that field sampling of basal ice often crosses discrete basal ice layers resulting in mixed $\delta^{18}O$ and δD values of ice formed from distinctly different δR_I compositions. Thus, the combination of natural variation in values of δR_I and the recovery of samples across discrete layers results in a scatter of $\delta^{18}O$ and δD values around poorly defined freezing slopes (Hubbard and Sharp, 1995). Therefore, Hubbard and Sharp (1995) propose the use of freezing envelopes where both δR_I and K can vary to include all possible

$\delta^{18}O$ and δD values of basal ice formed in closed and open meltwater reservoir system freezing events. The height of such an envelope is defined by the range of variation in values of δR_I along the Universal Precipitation Line, while the width of the envelope is dependent on the sensitivity of K included in individual basal ice samples (Hubbard and Sharp, 1995). A theoretical freezing envelope for a representative range of δR_I values at the Matanuska Glacier is given in Figure 4.

The co-isotpic freezing slopes (Jouzel and Souchez, 1982; Souchez and Jouzel, 1984) and envelopes (Hubbard and Sharp, 1995) previously described are designed for interpretations of basal ice samples formed in subglacial reservoirs with either no variation or well constrained variation in both the value of K and the composition of δR_I along the meteoric water line. In reality these forms of analysis may not be appropriate for dynamic open subglacial reservoirs where large volumes of water are moving through the system, which is characteristic of low elevation mountain valley glaciers such as the Matanuska. Due to the large volumetric flux of water relative to the rate of basal ice formation the value of F in this conceptual open system is many orders of magnitude less than I, subsequently requiring that O and I are essentially equal. The $\delta^{18}O$ and δD values of basal ice formed from discrete δR_I compositions in this system, therefore, will always correspond to values of K much less than 0.1. Thus, the production of basal ice will not measurably change the value of δR_I and the ice will reflect the $\delta^{18}O$ and δD

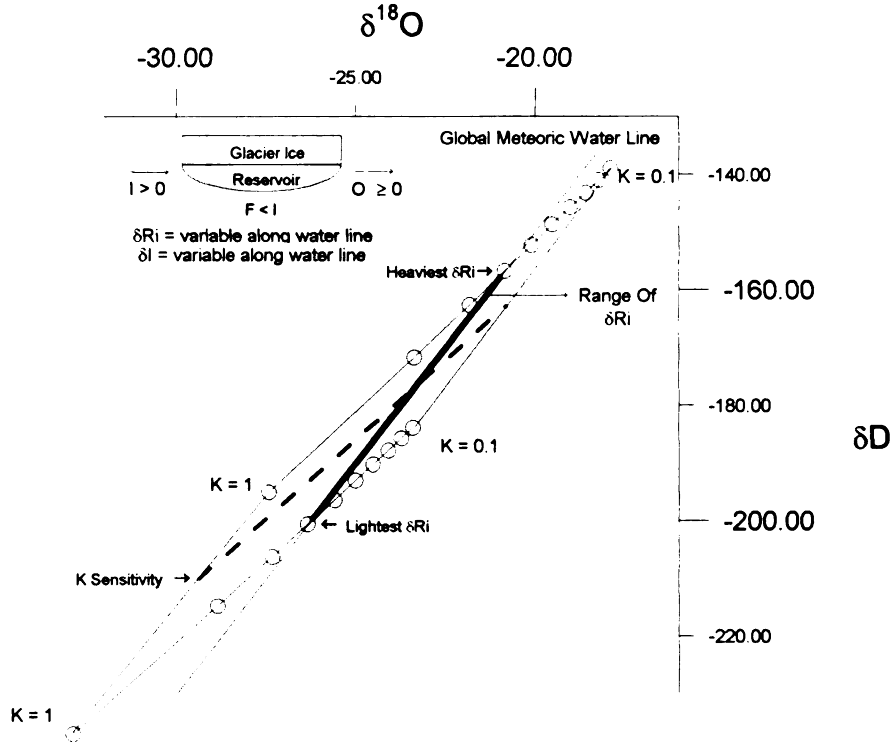

Figure 4. Freezing Envelope

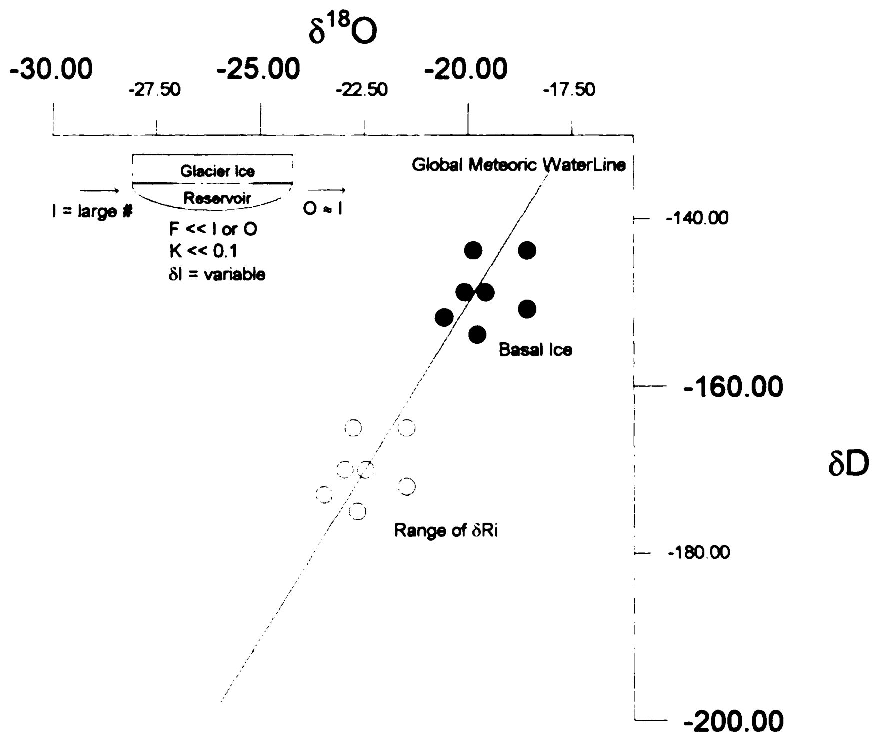

Figure 5. Open Meltwater/Precipitation Reservoir

value of O plus the maximum fractionation coefficients 2.91‰ and 21.2‰ for $\delta^{18}O$ and δD, respectively. Additionally, during discrete time intervals recent meteoric water (e.g., rain) may comprise a significant portion of the subglacial reservoir in low altitude glaciers. Contributions of precipitation to the subglacial reservoir may result in variation of δR_I both on and off the meteoric water line. The degree to which precipitation affects the δR_I of a reservoir is dependent on the volumetric contribution of precipitation and its $\delta^{18}O$ and δD values.

Therefore, we propose an additional model for the interpretation of the co-isotopic composition of basal ice formed in dynamic open subglacial systems. In this conceptual open meltwater/precipitation system, since the δ values of O and δR_I are equal, the values of $\delta^{18}O$ and δD for basal ice produced can be calculated from the equations:

$$\delta D = \delta_L D + 21.2‰ \qquad \textbf{(5)}$$

and

$$\delta^{18}O = \delta_L{}^{18}O + 2.19‰ \qquad \textbf{(6)}$$

where $\delta_L D$ and $\delta_L{}^{18}O$ are initial values of δR_I, and 21.2‰ and 2.91‰ are the isotopic enrichment of ice relative to δR_I (Lehmann and Siegenthaler, 1991). This form of analysis can be conducted irrespective of freezing slopes and allows for maximum variation in the values of δR_I, and thus basal ice values of $\delta^{18}O$ and δD, on and off the meteoric water line. A series of predicted basal ice $\delta^{18}O$ and δD values produced from a range of δR_I values, represented by vent discharge at the Matanuska Glacier, is given in Figure 5.

The presence of liquid precipitation in an open meltwater/precipitation subglacial reservoir allows for the incorporation of anthropegenic ^3H into the interpretation of basal ice formed in this system (Strasser et. al., 1996; Lawson et. al., in press). ^3H in basal ice produced by freezing subglacial water will be fractionated due to differences in spacing of vibrational levels of individual atoms in a water molecule. This spacing is proportional to the square root of the mass ratio of the heavy to light isotopic species. Thus, fractionation of $^3H/^1H$ is 1.4 times the fractionation of $D/^1H$, and the ratio of $D/^1H$ is 8 times that found in the mass ratio of $^{18}O/^{16}O$ (Strasser et al., 1996). In an open drainage system freezing event, the maximum fractionation of ^{18}O from the liquid to solid phase is 2.91‰ (Lehmann and Siegenthaler, 1991). Enrichment of ^3H, therefore, is approximately 4% from the liquid to solid phase (Strasser et al., 1996). Concentration variations in H^3 during single precipitation events and analytical error associated with lower detection limits are much greater than the 4% enrichment of ^3H due to freezing. Thus, ^3H concentration in basal ice, produced from variable subglacial reservoir concentrations of ^3H, will reflect the ^3H content of O.

FIELD SITE

The Matanuska Glacier is located in south-central Alaska where it flows northwest out of icefields in the Chugach Mountains. This large valley glacier is approximately 45km in length and ranges in width from 2.2km, at its source in the icefields, to 5km at the terminus (Figure 6). Over its 45km length, the glacier realizes a 3000m change in elevation from 3500m at the icefields to 500m at the terminus where, at it's present position, it has remained relatively stable for the last 200yrs (Williams and Ferrians, 1961). The northern portion of the terminus is stagnant and covered with debris, while the western portion is characterized by active ice and a lack of debris cover (Figure 7).

Subglacial hydrologic activity is most evident on the western portion of the terminus where ice flows through a small basin or overdeepening (Arcone et al., 1995). Along this margin numerous discharge vents can be observed in the form of upwellings or fountains. Many of these vents emanate from directly under basal ice exposures that range in thickness from 1m to more than 5m. Collective discharge from the vents coalesces into a proglacial catchment which subsequently discharges into a series of streams that flow west eventually forming the Matanuska River (Figure 7).

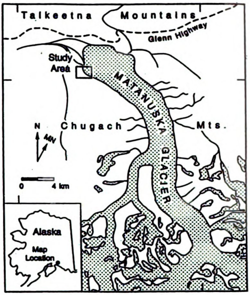

Figure 6. Regional Study Location

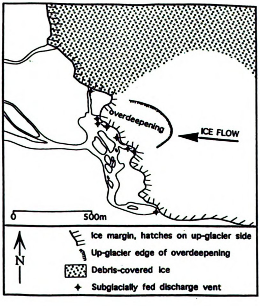

Figure 7. Local Study Locatrion

Methods

Discharge vents along the ice margin (Figure 7) were sampled because they represent the distal portion of the subglacial drainage system (Lawson, 1986; Strasser et al., 1992). This assertion is supported by observations of acute levels of suspended sediment in the vent discharge, which is characteristic of subglacially derived water and not water transported through clean englacial conduits. Vent samples were extracted either manually or through an ISSCO™ water sampler with a weighted sampling tube inserted into the vent throat. Samples were subsequently transferred to, and sealed in, 275ml Nalgene™ bottles for shipping.

Vent sampling was conducted on a diurnal interval based on the times of daily high and low discharge determined at a gauging station installed 300m down-stream from the vent catchment area (Figure 7). At this location, the stream is fed predominantly by discharge vents along with a small contribution from englacial ice-melt in the form of surface runoff from the glacier. The hydrograph (Figure 8) recorded at the gauging station shows a strong diurnal variation in stage, and thus subglacial discharge, in addition to a steady increase in flow during the spring months. Peak discharge occurs during June and July followed by a steady decrease in flow through the fall (Strasser, 1995; D.E.

Lawson unpublished data). Daily inspections of the gauging station hydrograph were made in order to adjust the sampling strategy to exact times of high and low discharge.

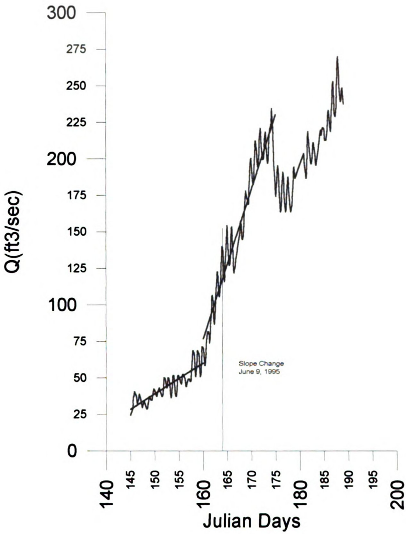

Figure 8. Discharge Hydrograph

Approximately 10mls of each vent sample was allocated for $^{18}O/^{16}O$ and $^{2}H/^{1}H$ ratio measurements. Sample splits were sent to Coastal Labs Inc. for analysis where the ratios of $^{2}H/^{1}H$ and

$^{18}O/^{16}O$ were determined using the standard calculation method:

$$\delta(‰) = \left| R_{(sample)} - R_{(standard)} / R_{(standard)} \right| * 1000 \qquad \textbf{(7)}$$

where R is equal to D/H and $^{18}O/^{16}O$ for the experimental sample or the laboratory standard respectively (e.g. SMOW). Analytical accuracy was reported as 0.1‰ and 1.0‰ for $\delta^{18}O$ and δD respectively.

Analysis for ^{3}H was conducted at the Michigan State University low level ^{3}H lab using a standard scintillation counting procedure (Kessler, 1988). Abundance of ^{3}H in a sample volume is expressed in ^{3}H units, where one TU is equivalent to one ^{3}H atoms to every $^{1}H*10^{18}$ atoms. Concentrations of ^{3}H in discharge have consistently been observed to be less than 20 TU at the Matanuska Glacier (Strasser et al., 1996; Lawson et al., in review). Low level ^{3}H analysis has been shown to be accurate to ±1 TU if electrolytic enrichment is applied to samples (Ostlund and Werner, 1962). 250mls of each sample were allocated for electrolytic enrichment and subsequent ^{3}H analysis.

RESULTS

The $\delta^{18}O$ and δD values measured in vent discharge samples collected from the Matanuska glacier after June 9, 1996 are presented in Figure 9. This time period is characterized by a rapid increase in meltwater production (Figure 8), which is a condition required for expansion of the basal drainage system and the subsequent accretion of basal ice (Alley et al., in press) Also included in Figure 9 are: Previously published $\delta^{18}O$ and δD values for basal ice samples collected in 1995 from site 95-1 near the glacier terminus (Strasser, 1996), average isotope values for basal ice (Strasser, 1996), and average isotope values for englacial ice based on an extensive unpublished data base of samples collected since 1978 from different locations near the terminus by D. E. Lawson (unpublished data). The local meteoric water line and the average value for precipitation in Figure 9 are also derived from the unpublished data base.

The vent discharge isotope values show a considerable amount of scatter about the local meteoric water line and vary from a maximum of -21‰ to a minimum of -24.6‰ with an average of -22.8‰ for $\delta^{18}O$, and from a maximum of -172‰ to a minimum of -189‰ with an average of -181.8‰ for δD. It is evident from the data that the average for vent discharge is more negative than average

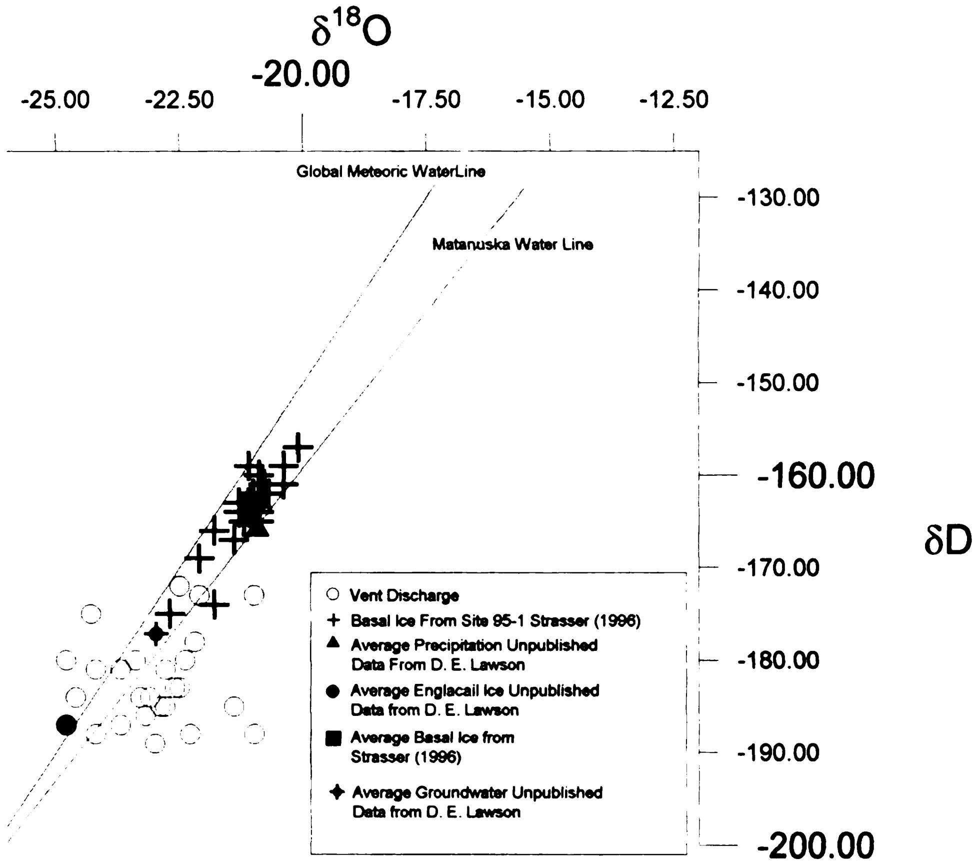

Figure 9. Stable Isotopes In Vent Discharge

basal ice or precipitation. Vent discharge, furthermore, is only slightly less negative with respect to the average $\delta^{18}O$ and δD values of englacial ice.

Vent discharge ^3H concentrations, much like $\delta^{18}O$ and δD, also show considerable variation (Figure 10). Vent discharge ^3H values were found to range from a maximum of 8.3 TU to a minimum of 2.3 TU with an average value of 4.6 TU. Compared to ^3H concentration measured in summer precipitation in Anchorage, Alaska (unpublished data from R. L. Snyder, U.S. Geological Survey), which is within 150Km of the Matanuska Glacier and generally experiences the same weather systems, it is apparent that vent discharge is diluted with respect to it's ^{3}H content. Conversely, discharge is enriched in ^{3}H relative to the concentrations of ^{3}H measured in englacial ice at the Matanuska Glacier (Figure 10) (Strasser et. al. 1996; unpublished data from D.E.Lawson).

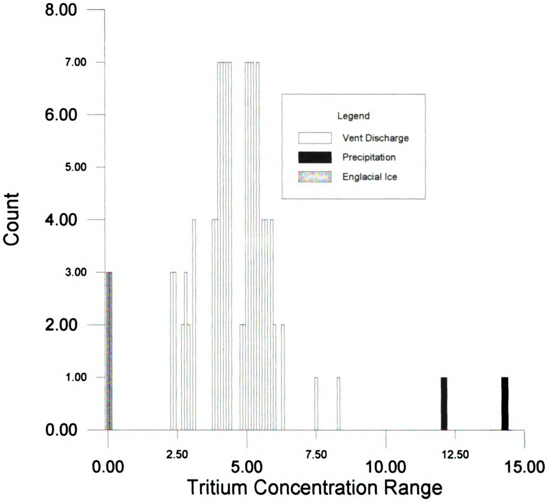

Figure 10. Tritium in Englacial Ice, Vent Discharge, and Precipitation

STABLE ISOTOPE DISCUSSION

Only the open meltwater/precipitation reservoir system freezing model is appropriate for interpretation of a genetic relationship between values of $\delta^{18}O$ and δD measured in vent discharge and basal ice at the Matanuska Glacier for two reasons: First, Alley et al. (in press) has calculated a gross basal ice accretion rate of $1.69 m yr^{-1}$ in the overdeepening at the Matanuska when the volumetric flux of water through the drainage system is $3.2 x 10^{6} m^{3} yr^{-1}$. Thus, in this system the value of K (e.g. fraction of the reservoir frozen) is orders of magnitude less than 0.1 requiring, therefore, that I and O are essentially equal and that the $\delta^{18}O$ and δD values of O and δR_I are equivalent. Second, given that vent discharge is comprised of a variable mix of englacial ice melt, groundwater, and recent liquid precipitation, Figure 11 shows that only the variation in $\delta^{18}O$ and δD values of precipitation are great enough to explain variation of these same isotopes in vent discharge (Strasser, 1995). This condition, in addition to the elevated levels of ^{3}H measured in vent discharge, suggests that recent liquid precipitation is present in significant volumes in the subglacial system and is responsible for the variation of $\delta^{18}O$ and δD values both on and off the water line in vent discharge.

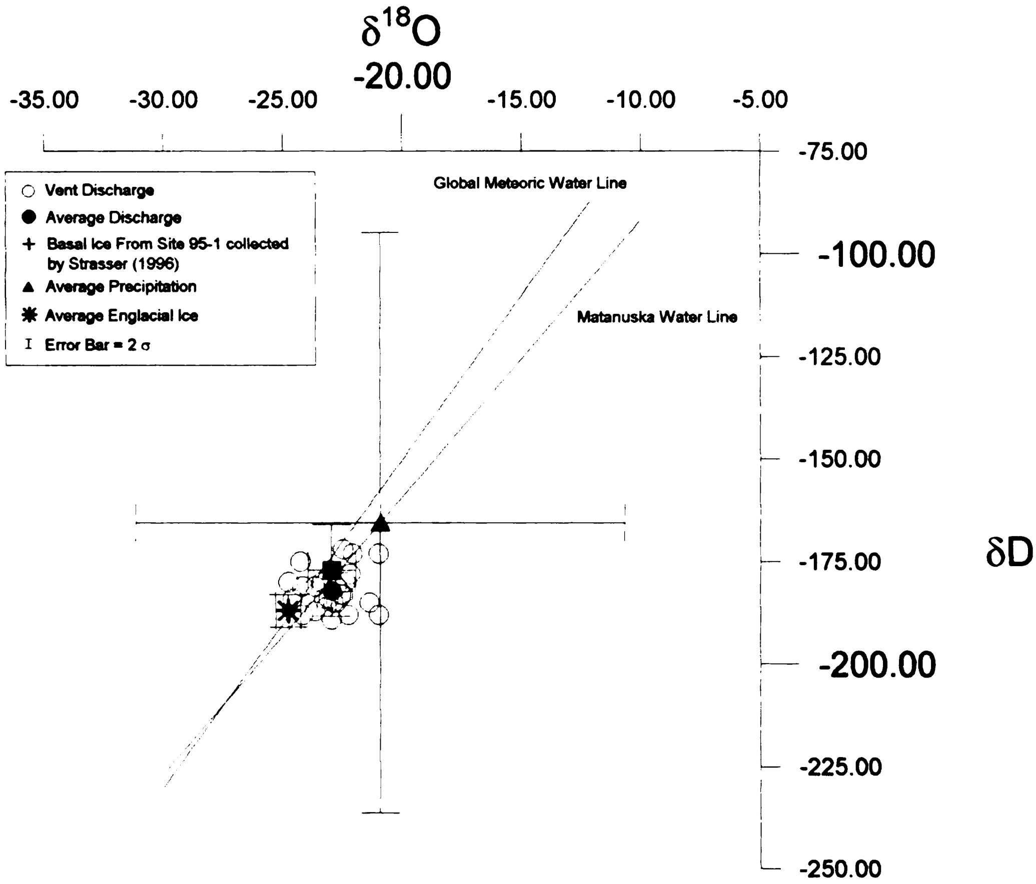

Figure 11. Influence Of Rain On Stable Isotopes In Vent Discharge

Application of equations 5 and 6 from the open meltwater/precipitation reservoir system model, to vent discharge samples, results in theoretical basal ice $\delta^{18}O$ and δD values that display a considerable degree of scatter about the Matanuska water line (Figure 12), and in this way are inconsistent with those values measured in basal ice by Strasser et al. (1996) at the Matanuska Glacier. It is instructive at this point to investigate the sampling method used by Strasser et al. (1996) to collect basal ice for ^{18}O, D and ^{3}H analysis. Core sections 5-10cm thick were collected and melted to provide adequate sample volumes for enriched ^{3}H measurements (Strasser et al., 1996). This process physically homogenizes the $\delta^{18}O$ and δD values of discrete basal ice horizons and therefore masks the variability of these values in the ice. Thus, the true $\delta^{18}O$ and δD range of discrete basal ice layers is not represented, and it is implicit that comparison of theoretical frozen discharge to basal ice is only reasonable if the theoretical values are averaged thereby representing a homogenized basal ice $\delta^{18}O$ and δD value produced from discharge (Figure 12).

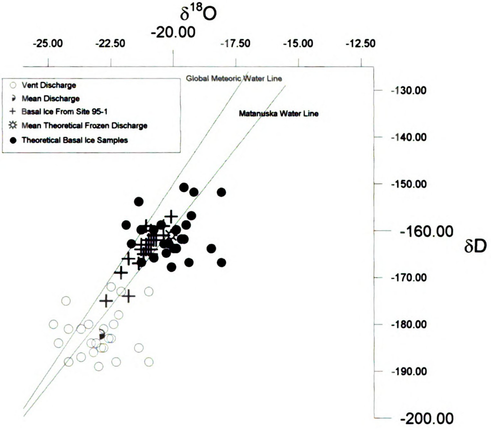

Figure 12. Theoretical Stable Isotopic Composition Of Ice Derived From Vent Discharge

Homogenization of basal ice $\delta^{18}O$ and δD values may also be caused by two additional mechanisms: First, if basal ice is forming in an open lattice framework (Strasser et al., 1996) and the δR_T of the reservoir is changing during that formation period, then basal ice $\delta^{18}O$ and δD values will reflect the averaged $\delta^{18}O$ and δD values of the two reservoirs. Second, Alley et al. (1996) theorizes that freeze-on and regelation may be occurring in the overdeepening either simultaneously or during different time periods during the year. Regelation through partial melting and subsequent refreezing of discrete basal ice layers may also homogenize $\delta^{18}O$ and δD values of basal ice.

Regardless of the homogenization effects of the sampling strategy, a strong linear trend is still evident in basal ice profile 95-1 collected by Strasser et al. (1996) (Figure 12). Simplistcally, all water at the Matanuska originated as precipitation and therefore high elevation samples will fall on the light end of the water line while low elevation samples will fall on the heavy end with respect to $\delta^{18}O$ and δD. If, then, subglacial discharge is a mixture of isotopically depleted high elevation precipitation (represented by englacial ice melt) and isotopically enriched lower elevation precipitation (e.g., rain), then discharge $\delta^{18}O$ and δD values will fall on the heavy or light end of the water line as a function of the relative contribution of either end member in a sample. Furthermore, if individual basal ice samples collected by Strasser et al. (1995) are approximately a seasons worth of ice production, then the $\delta^{18}O$ and δD values of those samples will represent the dominance of the heavy or light end-member in vent discharge over the course of a freeze-on season.

Using a one dimensional model Alley et al. (in press) calculated theoretical rates of basal ice accretion in an overdeepening similar to that of the Matanuska Glacier. Alley et al. (in press) calculates that when 10% of the Matanuska glacial discharge (e.g., 3.2×10^{6} $m^{3}yr^{-1}$) is routed through the overdeepening for a time interval of 0.1yrs, ~2mmy^{-1} of basal ice will be accreted. Net accretion rates of 16cmy^{-1} may be realized if 100% of the discharge is routed through the overdeepening for the same time period (Alley et al., in press). Given the aerial extent of the overdeepening at the Matanuska Glacier, and the density of discharge vents in that area (Figure 7), a reasonable estimate of discharge moving through the overdeepening may be 50% at any point in time. Following that assumption, a net accretion rate of ~8cmy^{-1} will be realized according to the Alley et al. (in press) one-dimensional model.

Theoretical accretion rates of basal ice can be used for time series analysis of variations in $\delta^{18}O$ and δD values when a reference datum in the ice can be identified. Strasser et al. (1995) labeled the ^{3}H peak in basal ice profile 95-1 as correlative with peak ^{3}H concentrations found in precipitation in 1963. Using the ^{3}H peak as a datum and an assumed annual accretion rate of 8cmy^{-1}, a total time of ~16yrs would be required to accrete the 127cm of basal ice below the ^{3}H peak in profile 95-1. Thus, the 26 samples collected below the ^{3}H peak datum would represent 0.62yrs of ice production each from 1963 to 1979.

A comparison of $\delta^{18}O$ values taken from basal ice profile 95-1 and precipitation deviation from normal, both as a function of time, is shown in Figure 13. The basal ice profile provided includes samples taken in the englacial ice as a means of graphically illustrating that there appears to be an isotopic dispersion zone between englacial ice and basal ice in the profile (Figure 13). An investigation as to the precise nature of possible dilution in this zone is beyond the scope of this paper. Contamination of $\delta^{18}O$ values in this dispersion zone precludes any reasonable comparison with precipitation deviation values, yet possible dilution of the peak ^{3}H value does not obfuscate its utility as a reference datum to 1963. A lower stratigaphic limit of the dispersion zone is included on the plot, and comparison of $\delta^{18}O$ deviation from normal to precipitation deviation is conducted only for samples beyond this limit.

The variation in $\delta^{18}O$ of basal ice samples and precipitation deviation as a function of time exhibits a series of peaks, in the positive direction, that are nearly in phase with each other (Figure 13). The $\delta^{18}O$ value of theorized basal ice from vent discharge for summer 1995, additionally, shows a strong positive deviation from normal, correlative with precipitation deviation from normal. Large negative deviations in precipitation, conversely, are not reflected in the basal ice $\delta^{18}O$ values, nor should they be. Generally the majority of subglacial discharge is comprised of ice melt, and therefore vent discharge $\delta^{18}O$ values are always at or near the lightest end member of the system,

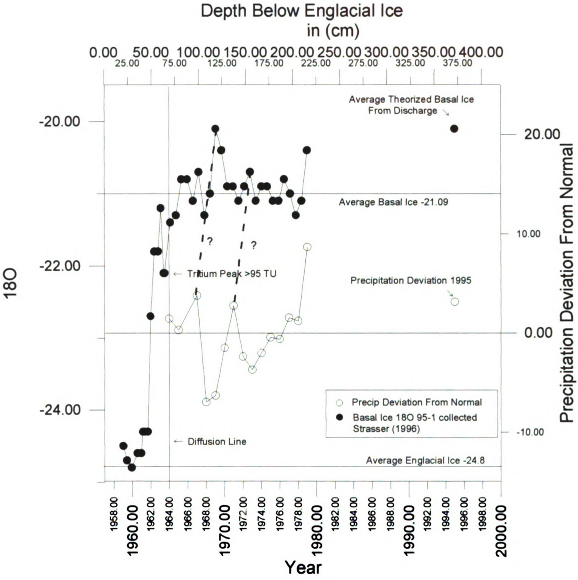

Figure 13. Relationship Between Precipitation Deviations And Variation In $\delta^{18}O$ Measured In Basal Ice

englacial ice. Only a large annual increase in the amount of low elevation precipitation (e.g., rain) can shift average annual

$\delta^{18}O$ values of vent discharge, and basal ice produced from vent discharge, away from the englacial ice end member. Thus, variation in the basal ice produced from discharge, regardless of the homogenization due to sampling, must reflect averaged variation in contributions of different end members in the subglacial reservoir. Oscillations in the relative amounts of precipitation present in vent discharge from year to year seems to be the only plausible explanation for the observed variation in $\delta^{18}O$.

It should be restated that accretion rates for basal ice samples presented in Figure 13 are assumed to be constant through time. Annual variation in subglacial hydrologic activity, due to a myriad of possible changes in atmospheric and subglacial conditions, require that accretion rates from year to year are not constant. Increased or decreased accretion rates between years could explain the peak phase shift in $\delta^{18}O$ observed in Figure 13. The sampling strategy of Strasser et al. (1996), alternatively, may have inadvertently crossed stratigraphic boundaries of basal ice produced during single freeze-on seasons resulting in homogenization of two or more years worth of basal ice production.

TRITIUM DISCUSSION

It is possible to approximate average seasonal ^{3}H content in discharge back to 1958 using the 1995 subglacial discharge ^{3}H concentration average of 4.6 TU and the historical record of ^{3}H concentration in precipitation for Anchorage AK (unpublished data from R.L. Snyder, U.S. Geological Survey). Long term summer month records (e.g. May-Sept.) of ^{3}H concentrations in precipitation for Anchorage AK. are presented in (Figure 14) (unpublished data from R.L. Snyder, U.S. Geological Survey). The 4.6 TU measured vent discharge average is 35% of the 13.2 TU average measured concentration in precipitation for Anchorage AK in the summer of 1995. Assuming this percentage represents an average volume of recent meteoric water in the subglacial system over the course of an ablation season, the range of ^{3}H in discharge from 1963 to 1995 corrected to a reference date of Dec. 31, 1995 would be 208.2 TU to 2.2 TU, respectively (Figure 14).

Estimating the range of ^{3}H in basal ice by this percentage method is a crude approximation. Strasser et al. (1996) reported a 6 TU measurement in subglacial discharge immediately following a precipitation event whose ^{3}H concentration was reported as 8 TU. A minimum value of 2.3 TU, which is close to background levels, was measured in discharge during the same summer over the course of a protracted period of dry sunny

Thus, a wide and variable range of 75% to nearly 0% meteoric water was present in the subglacial drainage system at any given time during the summer of 1995. Nevertheless, using the 75% meteoric water concentration in vent discharge as an average annual value, basal ice ^{3}H concentrations of >95 TU reported by Strasser et al (1996) could have only been formed from vent discharge during the time interval 1962-1966 (Figure 14). Based on the more reasonable estimate of 35% average meteoric water concentration in discharge, the >95 TU basal ice could have only been accreted during the years 1963 or 1964.

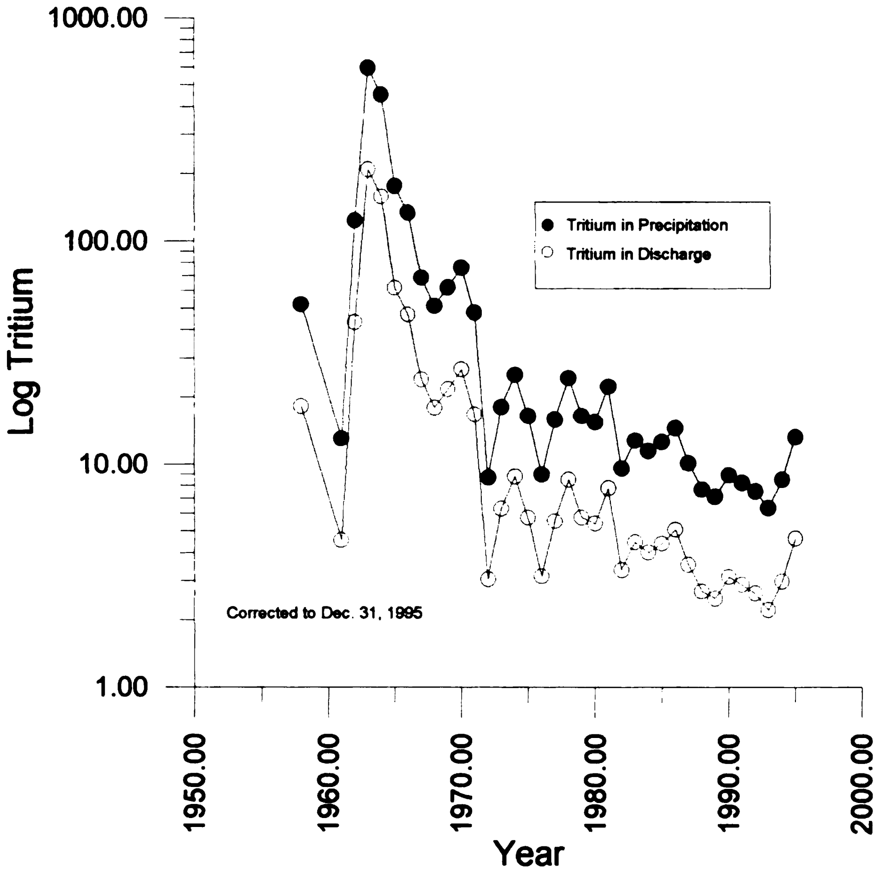

Figure 14. Simulated Concentration Of Tritium In Vent Discharge VS. Tritium In Precipitation(1995-1958)

CONCLUSIONS

The $\delta^{18}O$, δD and ^3H measured in vent discharge at the Matanuska Glacier Alaska are within the requisite ranges to account for variations in these same isotopes in basal ice. A simple open meltwater/precipitation reservoir fractionation model shows that the average theorized basal ice $\delta^{18}O$ and δD value produced from vent discharge is similar to those values measured in basal ice by Strasser (1996), and that there is a genetic relationship between vent discharge and basal ice. This result supports the hypothesis of Strasser et al. (1996) and Lawson et al. (in review) that basal ice is being accreted in a top down fashion in the overdeepening at the Matanuska Glacier from supercooled subglacial vent discharge. Alternatively, regelation may also be responsible for accretion of this basal ice. However, basal ice $\delta^{18}O$ values fractionated beyond 3‰ relative to $\delta^{18}O$ values of vent discharge, which is characteristic of a melting-refreezing process, have not been observed.

The range of $\delta^{18}O$ and δD values of theorized basal ice produced from vent discharge, furthermore, encompass all but the most negative samples from basal ice profile 95-1. The summer of 1995 was characterized as a season with greater than normal precipitation amounts, thus increasing the average volume of deviations in $\delta^{18}O$ values from basal ice with respect to deviations in precipitation, additionally, indicate further that basal ice formed in the overdeepening at the Matanuska Glacier is directly influenced by the percentage of recent meteoric water present in vent discharge. Depleted $\delta^{18}O$ and δD values of basal ice profile 95-1 were presumably accreted during dry years when less meteoric water existed in the subglacial drainage system. The sampling strategy of Strasser et al. (1996), however, homogenized discrete basal ice layers inadvertently obfuscating the true range and variability of ^{18}O, D and ^{3}H.

The 4.6 TU average measured in vent discharge demonstrates that an average of 35% meteoric water was present in vent discharge during the summer of 1995. Based on this percentage, which may be an over estimate due to the greater than normal precipitation volumes of summer 1995, the range of ^{3}H in basal ice produced from vent discharge through the years 1958-1995 would be 208.2 to 2.2 TU. These values are consistent with those measured in profile 95-1 collected by Strasser (1996) and indicate that the peak ^{3}H concentration measured in basal ice could only have been accreted during 1963 or 1964 from vent discharge.

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