Modeling Subglacial Hydrology in the Himalayas

by

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Abstract

The snowpack and glaciers of the Himalaya-Karakoram range feed several major river systems in Asia which provide water to over a billion people. Glacial retreat, glacial lake outburst flooding (GLOFs), surge behavior, and glacial ice mass balance are all likely strongly affected by subglacial hydrology. Unfortunately, little is known about Himalayan glaciers due to their remoteness and the danger of doing field work there. Recent advances in subglacial hydrological modeling may allow us to shed more light on subglacial processes that lead to changes in ice mass balance and glacial lake flooding. In this master's thesis, we present the first application of the SHAKTI subglacial hydrology model to a Himalayan glacier. We model the subglacial drainage network of Shishper Glacier, located in Gilgit-Baltistan, Pakistan, to understand its seasonal evolution and history of surges and GLOFs. Our results show that Shishper's subglacial system follows a similar seasonal pattern to past observed and modeled subglacial systems. We find that a central channel persists through the winter and serves as the basis for the subglacial drainage system throughout the melt season. We also investigate the 2017-2019 surge of Shishper Glacier and find that subglacial hydrology, while likely an important component of surging, cannot provide a standalone explanation for surges. This work serves as a nucleus for future subglacial hydrology modeling work in the Himalayas and provides a new framework for studying the effects of climate change on glacier dynamics, water availability, and glacier-related hazards in the Himalaya-Karakoram (H-K) region.

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Dedicated to my mom and my dad.

In loving remembrance of Gally (Galileo) Narayanan (2009-2023)

Om Shanti

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Contents

1 Glaciers in the Himalaya-Karakoram (HK) Region **11**

1.1 The Importance of Glaciers in High Mountain Asia 11

1.1.1 Climate and Weather Patterns 13

1.2 Glacial Retreat in the Himalaya-Karakoram 14

1.2.1 Previously Observed Glacial Retreat in the Eastern Himalaya 14

1.2.2 The Karakoram Anomaly 15

1.2.3 Predicting Glacial Mass Balance for the 21st Century 16

1.3 Chapter Conclusion 17

2 Water Scarcity, Governance, and Flooding in Pakistan **19**

2.1 Water Scarcity 20

2.1.1 Sources of Water in Pakistan 21

2.1.2 Water Governance 21

2.1.3 Indigenous Knowledge 25

2.2 Floods and Disasters 26

2.2.1 Monsoon-Related Flooding 27

2.2.2 Glacial Lake Outburst Floods (GLOF) 28

2.3 Gendered Impacts of Water Scarcity and Disasters 31

2.3.1 Sexual and Maternal Health 31

2.3.2 Gender Roles in Households and the Labor Market 32

2.3.3 Social Vulnerabilities and Strengths 33

2.4 Chapter Conclusion 34

3 Glacier-Related Hazards **37**
3.1 Anatomy of Mountain Glaciers 37
3.2 Surges 39
3.2.1 Suggested Mechanisms for Surging Behavior 39
3.2.2 Surges in the Karakoram 42
3.3 Glacial Lake Outburst Floods 43
3.3.1 Classification of Glacial Lakes 44
3.3.2 GLOFs in the Karakoram Region 46
3.4 Chapter Conclusion 47
4 A Primer on Subglacial Hydrology **49**
4.1 Subglacial Drainage Networks 50
4.1.1 Delivery of Meltwater to the Bed 50
4.1.2 Components of the Subglacial Drainage Network 51
4.1.3 Seasonal Evolution of the Drainage Network 57
4.1.4 Sliding Laws 59
4.1.5 Differences between the Polar and Alpine Environments 61
4.2 Observing Subglacial Hydrology 62
4.2.1 Remote Sensing Methods 62
4.2.2 In-Situ Methods 64
4.3 Modeling Subglacial Hydrology 67
4.3.1 SHAKTI: Subglacial Hydrology and Kinetic, Transient Interactions 68
4.4 Chapter Conclusion 72
5 Case Study: Modeling of Subglacial Hydrology in Gilgit-Baltistan, Pakistan **75**
5.1 Introduction 75
5.2 Model Setup 77
5.3 Preliminary Simulations 79
5.3.1 Winter Spin-Up 81

5.3.2 Discrete Meltwater Inputs . . . 83
5.4 Transient Simulations . . . 87
5.4.1 Melt and Precipitation Data Acquisition . . . 87
5.4.2 Model Setup and Parameters . . . 88
5.5 Chapter Conclusion . . . 89

6 Shishper Case Study: Results, Discussion, and Conclusions 95
6.1 Seasonal Evolution of the Subglacial Drainage System . . . 95
6.1.1 Temporal Visualizations of Seasonal Evolution . . . 96
6.1.2 Spatial Visualizations of the Subglacial Drainage System . . . 103
6.1.3 Surging and GLOFs between 2017-2021 . . . 111
6.1.4 Model Limitations . . . 115
6.2 Future Work . . . 116
6.2.1 General Improvements to Model Accuracy and Precision . . . 117
6.2.2 The Effects of Liquid Precipitation . . . 118
6.2.3 Proximal Lake Drainage . . . 119
6.2.4 Moulin (Supraglacial Lake) Drainage . . . 120
6.2.5 Coupled Hydrology and Ice Dynamics Simulations . . . 120
6.2.6 Contributions to Early Warning Systems . . . 121
6.3 Conclusion . . . 121

7 Supplementary Materials 123

Chapter 1

Glaciers in the Himalaya-Karakoram (HK) Region

1.1 The Importance of Glaciers in High Mountain Asia

High-mountain Asia (HMA) is often referred to as the "Third Pole" of the world because it contains the largest number and mass of glaciers outside of the polar regions [160, 112]. HMA is comprised of the Himalaya and Karakoram (H-K) ranges, which cover approximately 650,000 km^2 and 90,000 km^2 respectively [29]. Estimates of the number and volume of glaciers of the H-K vary widely. Estimates of the number of glaciers range from 30,000 to 95,000 [39]. Glacier volume is difficult to measure on large regional scales, but current estimates vary between 36000 to 51000 km^2 [19].

It is crucial to understand and predict runoff availability for the major economic activities in the downstream river basins, including hydropower and agriculture [29]. The Indus, Ganges, Brahmaputra, Amu Darya, Tarim, Yangtze, Yellow River basins all originate from the H-K, with meltwater supplying approximately 40% of river runoff [73, 127]. River basins originating in the Himalayas are shown in black borders in Figure 1-1. These rivers affect the lives of approximately 1.4 billion people, about 20% of the global population [73]. The glaciers play an crucial role in regulating the

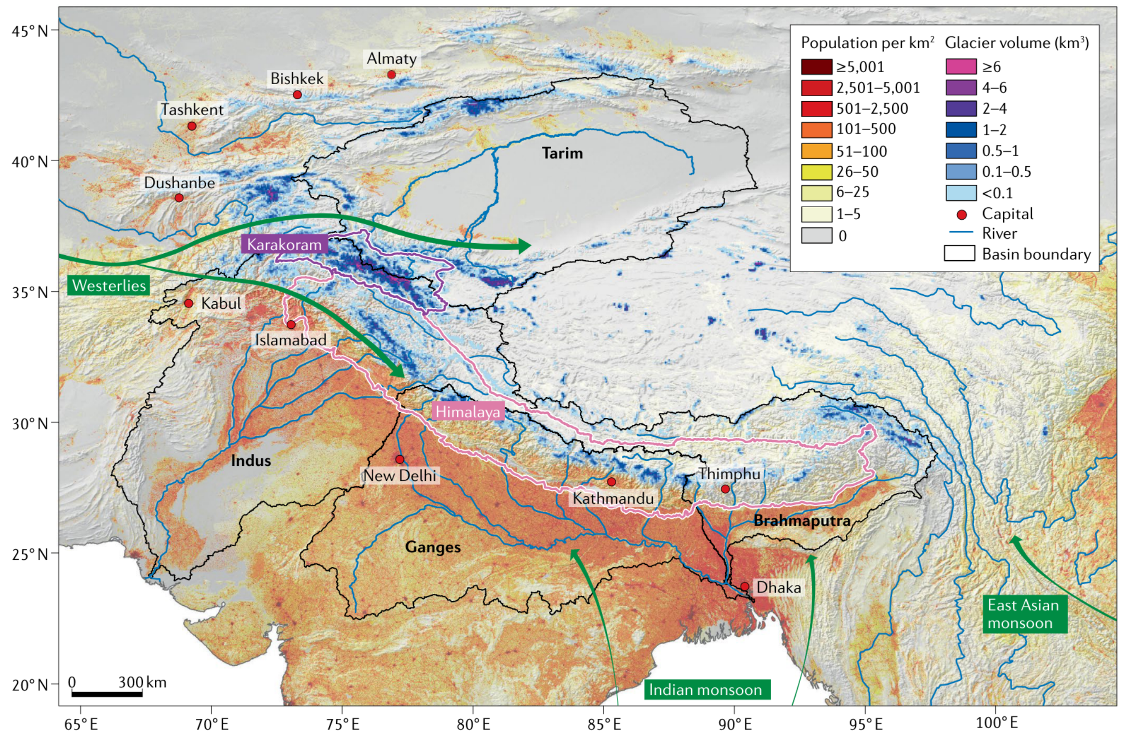

Figure 1-1: *Population per km^2 in the Indus, Tarim, Ganges, and Brahmaputra river basins, which are all sourced by glaciers in the Himalaya-Karakoram (the Yangtze and Yellow river basins are to the east of this map) [112].*

water supply of the region, acting as a buffer to the seasonal variations by providing meltwater runoff in the summer and fall [29]. Runoff volume is particularly sensitive to climate change because it depends on melting of seasonal snowpack and glacier ice [102]. Glacial meltwater runoff is particularly important for communities living in the river basins of the Karakoram and northwestern Himalaya, which receive less precipitation [29].

These river basins supply irrigation for agriculture in the region [73]. It is estimated that approximately 50% of India's utilizable surface water resources are provided by the Indus, Brahmaputra, and Ganges river basins, all of which are primarily sourced from the Himalayan glaciers [131]. Of all the rivers originating in the Himalayas, upstream discharge of the Indus and Brahmaputra basins are the most sensitive to climate change [73]; as a result, these areas will likely be heavily affected by water insecurity in the future due to their high population densities and dependence on river discharge for irrigation [73].

The hydrological system of the H-K is critically important for the hydroelectric power plants (HPPs) which are scattered throughout the mountains [112]. Glacier-fed basins in Nepal, India, and Pakistan particularly rely on hydropower because electricity shortages can be a severe issue there [112]. The volume of glacial meltwater corresponds directly with the hydroelectric power generation of the dams [119]. Because of this, power generation varies seasonally [112].

Most existing hydroelectric power plants have been built in the last 2-3 decades, with more dam projects projected to start as population and energy demand grow in Asia [112, 136]. However, natural hazards constitute a major danger to these expensive plants [136]. For example, glacial lake outburst floods (GLOFs) have destroyed power plants in the past, including the 45 megawatt Bhotekoshi hydropower plant in Nepal [40]. In addition, a recent trend has seen the placement of HPPs moving more and more upstream, putting them at greater risk of destruction or damage from GLOFs [136]. GLOFs can be triggered by natural events such as cloudbursts and flash floods [112]; for example, in 2013, a flash flood in the Upper Ganges in India led to a GLOF called the Kedarnath disaster [15]. This GLOF damaged a series of cascading HPPs downstream of the flood [15]. However, despite the increased risk of damage from natural disasters, population growth means that the H-K is forecasted to continue to expand on its hydroelectric power potential [112]. To safeguard South Asia's hydropower plants, it is important to understand where and when GLOFs may pose risks.

1.1.1 Climate and Weather Patterns

Glaciers may also play a key role in the regulation of weather patterns in the H-K region [29]. The influence of the Asian monsoon increases from west to east and from the north to the south in the Himalayas [29]. The interaction of the Asian Monsoon (AM) with westerly winds, also known as the Western Disturbance (WD) creates a complex hydrological cycle in the H-K mountains [30]. Glacier mass in monsoon-affected regions – i.e., the central and eastern Himalayas – is generally more sensitive to temperature changes because increases in temperature directly reduces snow accumulation and elongates the melt period [29]. In the central and eastern Himalayas, most glaciers are described as "summer-accumulation type" which gain mass primarily through snowfall due to the Asian Monsoon system [7]. In the eastern river basins, rain also contributes significantly to river flow. In the Karakoram region, however, most glaciers are classified as winter-accumulation types, meaning that they gain mass most during the winter due to storms caused by the Western Disturbances [29]. Because there is less precipitation in the Karakoram region, glacier meltwater and discharge through the Indus river system is the primary source of water for the region [12].

1.2 Glacial Retreat in the Himalaya-Karakoram

1.2.1 Previously Observed Glacial Retreat in the Eastern Himalaya

Studies have shown that since the mid-twentieth century, rising temperatures have led to unprecedented melt rates in the eastern Himalayas during the summer [112]. The negative mass balance in the Himalayas has also corresponded with increasing debris cover since the 1960s [19]. Shrinkage of glaciers has also led to fragmentation of large glaciers, which has increased the overall number of glaciers observed in H-K [87]. Both debris accumulation and glacier fragmentation can exacerbate melting, leading to a positive feedback loop of ice mass loss [112]. Figure 1-2 illustrates the decreasing glacier retreat from the eastern to western regions of the H-K mountain range.

From area change studies, [19] shows that despite slight regional variations in mass balance change, the central and eastern Himalayas have seen a continuous area shrinkage of -0.36% between 1960-2010. The loss of ice mass in the Himalayas has also corresponded to a rapid growth in various types of glacial lakes, a phenomenon which has major implications for human lives and infrastructure as well as hydrological

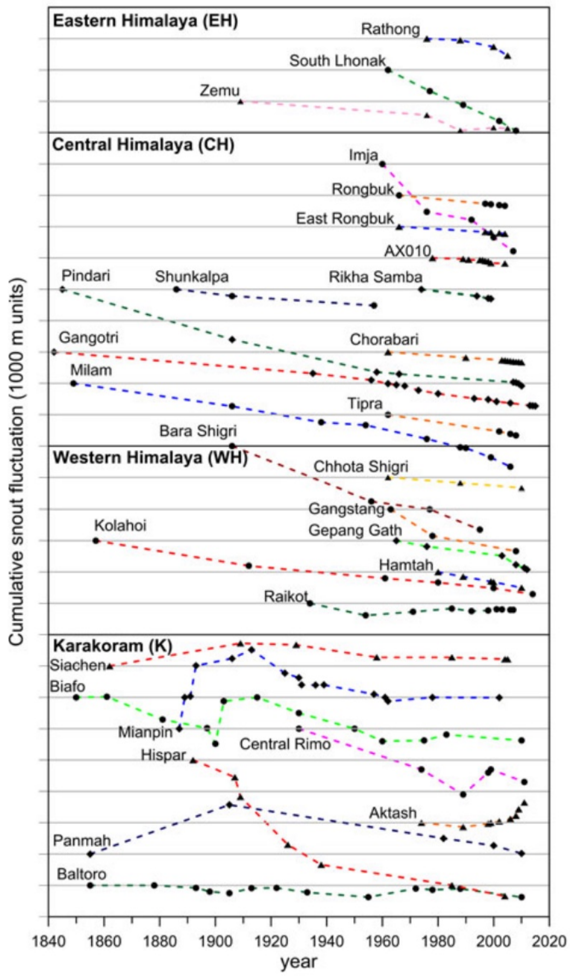

Figure 1-2: *Terminus positions in the Eastern, Central, and Western Himalayas show continuous glacial retreat through the 20th century. However, glaciers in the Karakoram have shown marginally increasing growth (Figure from [19]).*

systems and geomorphology [143].

1.2.2 The Karakoram Anomaly

In the past few decades, the glaciers of the Karakoram region showed an unusual trend of stable or even marginally increasing mass balance [88], a phenomenon that is still not well understood. This is commonly referred to as the Karakoram Anomaly [97]. Rather than retreating, Karakoram glacier fronts have been observed to exhibit stability or even advance in past decades [65]. In the Karakoram region, summer temperatures decreased since the 1960s, a trend that contrasts with the increasing summer temperatures in the eastern Himalayas [55]. This may have explained the anomalous thickening and lengthening of glaciers in the Karakoram Himalaya, as opposed to the warming and thinning that is seen in other glaciers worldwide; however, the underlying factors behind these decreasing summer temperatures are not well understood.

1.2.3 Predicting Glacial Mass Balance for the 21st Century

Alpine glaciers worldwide have consistently exhibited major retreat and lost hundreds of gigatons of mass in the past two decades alone [69]. It has been observed that mass loss in the Himalayas has seen a steady increase since the mid-19th century [95, 84]. Similarly, indicators show that temperatures have been steadily rising across the entirety of the Himalayas since at least 1959 [159].

A very recent report from the International Centre for Integrated Mountain Development (ICIMOD) found that between the years of 2010-2019, the Karakoram Anomaly is reversed and joined the rest of the world's glaciers in retreat. This is probably due to a recent increase in summer temperatures in the region [75]. According to the report, H-K glaciers are predicted to lose 30-50% of their mass by 2100 if warming is restricted to 1.5º to 2ºC [75]. This percentage increases if warming exceeds 2ºC. The number and size of glacial lakes will also increase drastically, causing more severe GLOFs [75]. This has very important implications for the future of water availability: "peak melt" is expected to occur by 2050, after which time glacier runoff will decrease [110]. This will affect water availability in all of the glacier-fed river basins in Asia, especially in the Indus, Ganges, and Brahmaputra basins where a combined 129 million farmers currently depend on glacial runoff to irrigate their crops [110]. The effects of this mass loss are wide-ranging, from reductions in permafrost and groundwater to increased landslides, avalanches, and flooding, as well as consequences for natural ecosystems and biodiversity [110].

1.3 Chapter Conclusion

It is difficult to overstate the wide-ranging importance of the glaciers in the Himalaya-Karakoram (H-K) range. Often referred to as the "Third Pole" due to the immense amount of ice sequestered there, the H-K range supplies meltwater runoff that feeds numerous river basins throughout Asia. Over a billion people rely on these river basins for agriculture, hydropower, climate regulation, and more. Since at least 1950, these glaciers have been observed to shrink at an accelerated pace due to climate change. This trend is only going to continue through the next century as temperatures and melt rates increase. In order to ensure South Asia's success, strategies must be developed to increase the sustainability of its water resources.

Chapter 2

Water Scarcity, Governance, and Flooding in Pakistan

NOTE: The purpose of this chapter is to contextualize the technical aspects of Himalayan glaciology within its broader humanitarian impacts. We have attempted to provide an impartial summary of the social and political issues surrounding Himalayan glaciers in the Karakoram range. This summary is very abbreviated and does not encompass many of the nuances that are required to fully understand the issues. For more information, please see the referenced works.

As climate change advances, the future of water supply in the Karakoram Range grows dire. Glaciers are receding everywhere worldwide, including in the Karakoram [110, 73]. In particular, Pakistan's water supply, which comes from the Indus Basin, relies heavily on water from glacial water towers because of its naturally drier climate and large agricultural economy [112]. The population of Pakistan is also growing rapidly, putting strain on its agricultural resources. Food security is a particularly important issue in Pakistan; therefore, managing its water resources is a critical need for Pakistan's political and socioeconomic growth [117].

In addition to the potential for food insecurity due to shrinking glaciers in the Karakoram, it has been shown that glacial lake outburst floods (GLOFs) and glacial surge behavior occur with increased frequency with rising temperatures [94, 22]. This has especially affected rural highland populations in northern Pakistan. As glaciers shrink and termini retreat, glacial lakes grow in size and risk potential (since larger lakes can cause more destruction to areas downstream) [94]. This creates a myriad of hazards that are important for lawmakers, scientists, and local citizens to consider. Furthermore, GLOFs have devastating effects on the lives of the people living in close proximity to the glaciers. Protecting the needs of these people presents a significant challenge for climate change adaptation.

2.1 Water Scarcity

Although Pakistan contains the largest number of glaciers outside the poles [77], it is one of the most water-stressed countries in the world. The presence of larger state and colonial governments has had a direct negative impact on water availability. Traditional water systems have been neglected in favor of “modern” or Western infrastructure, which is not as well-suited to conditions in South Asia [137]. The consequence of this is that much of the H-K region, including much of Pakistan, northern India, and Nepal now suffer from water shortages [137]. More than 80% of Pakistanis now face extreme water scarcity for at least one month out of the year [106]. In the next five years, Pakistan will reach "absolute" water scarcity [32]. About 92% of Pakistan’s land is classified as arid or semi-arid; however, being a primarily agrarian economy, 97% of Pakistan’s water is used for agriculture while the remaining 3% is used for domestic and industrial purposes [151]. In addition, its population is growing very rapidly, calling not just future water availability, but also food availability, into question [117]. As per-person water availability decreases, it is important to understand the reasons why Pakistan is facing such challenges and to implement better infrastructure for ensuring its water security in the future.

2.1.1 Sources of Water in Pakistan

The rivers in Pakistan are mostly glacier-fed [117]. As climate change advances, the supply of meltwater from Karakoram glaciers into the Indus basin are expected to increase until 2050 as glacial melting accelerates [117]. However, after 2050, as the glaciers shrink, glacial meltwater volume will steadily decline. Warmer temperatures will lead to increased evapotranspiration and decreased soil moisture, which will require more irrigation water [117].

Precipitation also amounts for a large percentage of the water supply. Monsoon rains, which occur between July and August, account for 70% of Pakistan's total water [1]. However, as we will discuss next, Pakistan's ability to store water is quite poor, so much of this monsoon rain supply is not able to be stored.

2.1.2 Water Governance

Some Pakistani scholars claim that a great deal of Pakistan's water stress can partially be blamed on government mismanagement leading to crumbling infrastructure [11]. One scholar even claims that the government has taken a "callous" approach to water management [101]. In particular, there is little regulation of irrigation practices, and agricultural practices are inefficient [101]. The country's main crops are sugarcane, rice, wheat, and cotton [151]; both sugarcane and rice are very water-intensive crops to grow, and are thus not practical for the dry climate. As a result of these inefficient practices and water-intensive crops, the water column is significantly lowered and underground aquifers are severely depleted [11].

Additionally, Pakistan's water storage capacity is diminished; its reservoirs are only able to store 30 days worth of water supply, whereas other countries such as India have up to a 220 day capacity [11]. This means that even if there is ample rain, it is not able to be stored, reducing the country's resilience to changes in weather. In addition, conflicts between the states of Sindh and Punjab have led to tensions about partitioning water, which has led to slowdowns in passing public policy [79].

However, a great deal of Pakistan's issues in water management come from colonial-era changes. For example, the British, who had colonized current-day Pakistan and India in the 1800s, built a large canal system drawing water away from the Indus River, accompanied by a large rail network, for the purpose of growing and exporting cash crops more easily [101]. To accommodate this canal system and railway, they underwent a massive deforestation project [101]. The continuing deforestation of Pakistan, which follows colonial precedent, has greatly exacerbated the effects of climate change, as forests act as important buffers for flood events [13]. Geographers also note that the colonial creation of these "Western-style" irrigation system destroyed the natural cycle of high-frequency, low-intensity floods which nourished Pakistan's native ecosystems [101]. Now, floods occur at less frequency but higher intensity, causing major destruction and human displacement [101].

Pakistan-India Water Sharing

The border between current-day India and Pakistan was drawn by the British in the 1940s, but failed to take rivers and already-existing irrigation systems into account [78, 101]. As a result, all of Pakistan's rivers flow first through India, reducing Pakistan's sovereignty and ability to manage its own water resources (Figure 2-1). Government officials and scholars in Pakistan are especially concerned with the large number of existing and planned hydropower projects on the Indian side of these rivers, which can result in sudden releases of water or stoppages of flow to the Pakistani side [79, 151, 78]. However, due to the deteriorated state of diplomatic relations between Pakistan and India, it has been difficult for the two countries to work out a fair and equitable solution for sharing water resources [79].

The Indus Waters Treaty (IWT) of 1960 was the first piece of water legislation between India and Pakistan. It has been regarded as one of the biggest success stories in water diplomacy, as it has survived through several wars between India and Pakistan [78]. Through this treaty, the rights to the Indus, Jhelum, and Chenab rivers were given to Pakistan, and accordingly the Ravi, Beas, and Sutlej to India (Figure 2-1) [78]. There are many scholarly opinions as to whether or not the IWT was "effective" or fair; certainly, it worked in that it allowed each country to serve

Figure 2-1: *All of Pakistan's critical rivers flow first through India. This poses a major challenge for joint watershed governance between Pakistan and India [78]*

its own self-interest (barring the fact that the three rivers "given" to Pakistan first flow through India), but a major flaw of the IWT is that it did not work out a way for the two countries to jointly manage the Indus watershed [8]. Also, China and Afghanistan, which also share the Indus basin, were not part of this treaty [8]. Now, there is a possibility that Afghanistan may try to build dams on the Kabul River, which also flows into Pakistan; this has the potential to lead to further restrictions of water flow [78].

As climate change causes increased water stress across the basin, the division of the Indus basin across international borders has severe implications not just for water security, but also for disaster management [8]. For example, a downpour in 2014 in the disputed Indian state of Jammu and Kashmir led to massive losses of life both in India and Pakistan. The lack of a joint flood management plan exacerbated the losses of lives in both countries [8]. Natural disasters such as cyclones, avalanches, and GLOFs are already increasing with climate change. As a result, there has been a recent push by climate researchers on both sides for India and Pakistan to coordinate data-sharing and disaster relief planning [123].

Inter-National Governance in South Asia

Effective water governance is key to improving water availability and implementing sustainable water management practices in the Himalaya-Karakoram region. Currently, there is a lack of effective bilateral and multilateral agreements between the countries that share water in the H-K region, mostly owing to tensions and disagreements between governments in South Asia [142]. For example, although India and Bhutan have an effective water management agreement through which both countries benefit, the agreements between India and Bangladesh, Nepal, and Pakistan are much less balanced due to mutual mistrust and differences in both power and attitudes toward water governance [142]. In particular, because upstream water management so profoundly impacts downstream water quality and availability, inter-national agreements are necessary to ensure equitable sharing of water [142].

Although many small-scale technical solutions to water shortages have been proposed and implemented (e.g., flood walls or improved irrigation systems), the primary macro-scale solution advocated by water experts studying the H-K region is the river-basin approach [142]. Managing resources at the river-basin scale allows maximum benefit from infrastructure projects and minimizes the risk of water-related hazards [137]. In particular, the International Centre for Integrated Mountain Development (ICIMOD) lays forward a detailed plan called "Multiscale Integrated River Basin Management" [109]. The plan integrates a variety of social, economic, and environmental factors and joins together multiple sectors affected by water use including domestic water use, agricultural irrigation, and hydropower [109]. In the past, a similar approach has been taken in other places to a great degree of success; for example, in Colombia, partnerships between conservation/advocacy groups and various community-based organizations led to improved economic status and increased fertility of the area surrounding the La Cocha Lagoon which was being degraded due to forest over-exploitation [109]. An approach like this offers a rare hope to improve the lives and prospects of billions of people in South Asia; however, poorly drawn colonial-era borders and political tensions in South Asia make the implementation of the river-basin approach unlikely.

2.1.3 Indigenous Knowledge

Understanding traditional and Indigenous knowledge is an essential tenet of climate change studies [6]. It is well-known that Indigenous and traditional knowledge is one of the best tools for understanding local geology and ecology and adapting to climate-related hazards [116]. Indigenous people often promote sustainable stewardship of land, and preservation of natural and cultural resources [116]. South Asia, which is home to the Himalayas, has a long and arduous history with colonization, exploitation, and Westernization. This has led to a myriad of land use changes that exacerbate water scarcity and flooding. It has also led to the unfortunate loss of a large body of Indigenous knowledge which had been passed on for generations. Thus, it is of critical importance for people who practice Western science to understand and incorporate remaining Indigenous knowledge into their methodologies going in the future.

Glacier Grafting

Glacier grafting is a traditional practice performed in Gilgit-Baltistan, Pakistan, the region in which this thesis' study area is located (see Chapter 5). The tradition was first mentioned in ancient texts in the 1300s C.E., and was also documented by the British in the early 1800s [48]. It is used to encourage the growth of glacier-like ice patches [104], which supply meltwater that can be used for agricultural irrigation. In Gilgit-Baltistan, scholars estimate that the technique of glacier grafting has the potential to transform 90,000 hectares of land per district into arable land [104].

The local folklore holds that glaciers are living entities that can be identified either as male or female [48]. Male glaciers are generally debris-covered and gray in color, whereas female glaciers are blue or white [48]. Male glaciers generally move more slowly and generate less meltwater, whereas female glaciers move more quickly and generate more meltwater [48]. To graft or "breed" the glaciers, local people take 300-kilogram chunks from a male glacier and a female glacier and combine them with Indus River water. The mixture is transported to a pre-decided location using insulated packs [48]. The pre-decided locations are generally in shadowed caves or deep pits high in the mountain (between 4000-5000m), where temperatures remain below freezing year-round [104]. The people then cover the ice with ash and mud and seal the hole with large stones. After holding festivities and rituals associated with this "glacier wedding," local villagers leave the glacier alone. Within 10-12 years, they can expect a new "glacier" to form [48].

Recent glacier grafts in Gilgit-Baltistan have yielded an 80% success rate [48]. This is no surprise; by placing a critical mass of ice in a suitable environment, it will be able accumulate enough precipitation such that accumulation exceeds ablation, leading to growth [48]. Villagers from the Shigar district of Gilgit-Baltistan said in an interview that plenty of water had been flowing continuously since performing a glacier graft, which allowed them to reliably cultivate wheat, barley, and vegetables in areas that had seen only intermittent flow before [48].

In the face of climate change, glaciers in Gilgit-Baltistan and all over the world are expected to shrink, reducing the amount of meltwater they can generate for food production. However, traditional practices like glacier grafting provide encouraging solutions for climate change adaptation.

2.2 Floods and Disasters

Water security is one side of the coin of Pakistan's water crisis; flooding is the opposite face. Both are phenomena exacerbated by climate change, and both are extremely dangerous. Other hazards such as avalanches and landslides are also becoming increasingly common as glaciers become less stable and permafrost melts. Further, the Indus basin is one of the most affected by extreme weather events; of all the basins in the H-K, it saw the most number of people killed between 1980 and 2015 [9].

2.2.1 Monsoon-Related Flooding

Large-scale, devastating flood events occurred in Pakistan most recently in August 2022, causing thousands of deaths, billions of dollars in damage, and temporarily dominating the global news cycle. Although the floods held the world's attention for only about a week, the severe repercussions of the floods will continue for months and years.

A majority of Pakistan's rainwater is delivered between July and August, corresponding with the Asian monsoon cycle. In 2022, however, Pakistan received about 190% of its yearly average rainfall [152]. The extremity of this flooding event is thought to have been exacerbated by the prolonged, severe heatwaves that affected Pakistan from May to June, creating a low-pressure system that caused the amplification of the existing monsoon rain system [107]. The system of embankments and barrages initially built by the British in the 1800s, originally meant to increase farmable land for cash crop exports, exacerbated the floods even further [32]. These structures prevented excess water from flowing back into the Indus River system, causing the floodwater to accumulate and harbor waterborne diseases including malaria [32].

During this time, about 1/3 of the country was completely submerged. 15,000 people were left dead or injured, 2 million homes were destroyed, and 4 million acres of agricultural land were severely damaged [1]. The cost to rebuild from these floods is over 30 billion US dollars, a cost unbearable by Pakistan by itself [115]. Although it received approximately $10b in international aid, Pakistan is still reeling from the damage caused by the 2022 floods [1].

These types of flood events are expected to increase with more and more frequency throughout the coming decades [107]. Returning to more traditional irrigation systems may alleviate the severity of floods in the future, but it will be difficult to make such a large-scale economic transition. Combined with glacier-related floods and disasters, monsoon-related floods put the entirety of Pakistan at risk.

2.2.2 Glacial Lake Outburst Floods (GLOF)

GLOFs occur when glacial lakes burst suddenly out of their natural impoundments, causing large-scale destruction of lives, infrastructure, and crops downstream. GLOFs occur all over the world wherever there are glaciers. In the Karakoram range, a higher frequency of glacial surge activity has also been associated with a greater risk of GLOFs (Chapter 3) [22]. However, the Himalaya-Karakoram and the Andes ranges have received the largest amount of scientific attention due to the large, vulnerable human populations downstream of these floods.

Although lake drainages can vary greatly in severity, the devastation caused by GLOFs is hard to understate. One of the most infamous recent GLOFs in South Asia, the Kedarnath Disaster, occurred in 2013 [5]. Following a cloudburst flash flood, Chorabari Lake in Uttarakhand, India was inundated and caused its moraine dam to fail [5]. The death toll was placed at about 6,000 people, including local residents of the town of Kedarnath as well as religious Hindu and Sikh pilgrims. The cost of rebuilding the bridges, houses, and damaged hydropower plants was hundreds of millions of USD [5].

Smaller GLOFs occur quite commonly all over the Himalayas. In northern Pakistan alone, there are 36 glacial lakes that are in danger of flooding [22]. Shisper Glacier is just one of them. In May 2022, following the hottest spring in Pakistan since 1961, the glacial lake at Shishper's terminus drained suddenly, discharging at 10,000 cubic feet per second [118]. The GLOF destroyed agricultural land, homes, and two power plants downstream, leaving much of the surrounding Hunza Valley without electricity [118].

In this area, lake drainages commonly cause damage to agricultural lands and people's livelihoods, leading to economic losses and emotional and physical trauma. People near the GLOF-prone lakes live in a state of continuous mental stress, and paramedics have observed widespread anxiety, depression, and post-traumatic stress disorder (PTSD) amongst villagers [46]. According to the film "The Sky is Far, the

Earth is Tough" by Pakistani filmmaker Haya Fatima Iqbal, opium is commonly used to alleviate the grief and pain that people feel [74]. One woman affected by a GLOF in a nearby valley said that her husband smoked opium all day, which caused her to take on double her usual household tasks [74]. Gilgit-Baltistan has also seen an increased suicide rate in young people, especially women [46].

Early Warning Systems

Early warning systems (EWSs) can alert local people of GLOF-related danger and prevent losses of life. At Shishper Glacier, an early warning system has proven effective at protecting people nearby from outburst flooding. The Pakistan Meteorological Department (PMD) regularly monitors Shishper due to its close proximity to human settlements and to the Karakoram Highway [76]. The PMD uses satellite data collected every 16 days, but these measurements are constrained by whether or not there is cloud cover over the area [76]. In addition, a live camera installed at the glacier can help assess the lake's water level [76]. Other EWSs in the Himalayas have made use of flood water routing models and flow hydrographs to assess which areas will be inundated and when [89]. In some areas, young men trek up to the glacier and listen for loud sounds that might signify an avalanche or landslide that may trigger a GLOF, which can reach downstream communities within minutes. They then use radio phones to alert communities downstream [74].

Although EWSs have been made possible through a combination of satellite imagery and in-situ measurements, none yet have the certainty or ability to trigger sirens or automatically alert nearby people of danger [89, 76]. All current EWSs require specific and constant attention from scientists in order to work properly. As the number of dangerous glacial lakes near human settlements increases in the Himalayas, it will become even more difficult to dedicate enough attention to each dangerous glacial lake. Therefore, it is of great importance to fully automate these EWSs and to make them more accurate so that evacuations and prevention techniques can happen on time.

GLOF Prevention: Lake Drainage

GLOF prevention measures can go a long way in preserving lives and infrastructure. A famous example of GLOF prevention is the lake drainage of Lake Imja in Nepal. In 2016, Imja Lake near Mt. Everest was assessed to be in danger of outburst flooding. The lake, which was 149m deep in some places, had been destabilized by recent earthquakes [145]. Army personnel and sherpas took 6 months to build a drainage system that lowered Imja Lake's water level by 3.4 meters, slowly releasing about 4 million cubic meters of water over the course of two months [145]. By doing this, the risk of GLOF was significantly reduced. This is a promising method for draining especially dangerous glacial lakes; however, the money, time, and specialized personnel required to construct an outlet and drain glacial lakes is extremely intensive and often infeasible for most sites.

United Nations Development Programme: GLOF-I and GLOF-II

The United Nations Development Programme (UNDP) has brought special attention to GLOFs in High Mountain Asia through the GLOF-I and GLOF-II projects. GLOF-I, "Reducing Risks and Vulnerabilities from GLOF Northern Pakistan," conducted between 2011-2016, aimed to establish an institutional framework and lay the groundwork for building a community-based response and adaptation program for the Gilgit-Baltistan and Khyber Pakhtunkhwa areas in Pakistan [4].

GLOF-II "Scaling Up of Glacier Lake Outburst Flood (GLOF) Risk Reduction in Northern Pakistan," which is currently ongoing, is a scaling up of the GLOF-I project. The project has received nearly $40 million in funding and is conducted in partnership with the Pakistan Ministry of Climate Change [3]. The project ranges in goals, from installing more automated weather stations (AWSs) and discharge measurement apparatuses to increasing the resilience of mountain communities. The project involves implementation of new irrigation and slope stabilization schemes, installation of small-scale infrastructure, and the establishment of Community Based Disaster Risk Management Centers (CBDRMCs) [3]. In particular, gabion walls control erosion, provide retaining structures, and help to reduce the velocity of bursting water [3]. Sea buckthorn, a deciduous shrub native to Asia and Europe, is being widely planted to prevent erosion and stabilize slopes at risk of landslides [3]. At the study site for this thesis work (see Chapter 5), Shishper Glacier, many efforts are in place including the improvement of irrigation systems and the planting of alfalfa and poplar cuttings to improve slope stability [3].

A central feature of GLOF-II is conducting trainings targeted towards community members in GLOF-prone areas [4]. The GLOF-II project has conducted mock drills and community awareness sessions for people in target areas. Further trainings aim to improve journalists' knowledge of disaster risk preparedness and to introduce concepts of community resilience.

2.3 Gendered Impacts of Water Scarcity and Disasters

The impacts of natural hazards such as GLOFs have a particularly severe impact on women, particularly poor women. As a result, many development programs have focused specially on including women's needs and utilizing their strengths for climate adaptation.

2.3.1 Sexual and Maternal Health

In the recent monsoon-related floods of 2022, gender-based violence and poor maternal outcomes reached an all-time high in Pakistan [56]. Much of this is due to the destruction of support systems such as domestic violence shelters and hospitals where women previously obtained maternity care [56]. In particular, displaced young women and girls are subjected to sexual violence when they are separated from their male family members who are traditionally responsible for their safety – until they reach the safety of a shelter, they often risk being coerced into sex or falling into the hands of human traffickers [56]. New mothers also have been observed to stop breastfeeding due to the stress of displacement and lack of privacy in the shelters, leading to additional poor outcomes for them and their children [56]. The lack of access to hospitals and usual maternal care during these disasters also puts expectant and new mothers and their children at risk [56].

2.3.2 Gender Roles in Households and the Labor Market

Traditional household dynamics, especially in high-mountain rural communities, often put women at higher risk compared to their male counterparts. Due to the highly gendered nature of the labor market in Pakistan, women are the first to be let go from jobs during market shocks [115]. In addition, when a household faces monetary stress, women tend to enter the labor market, which increases their total amount of work including household chores [115]. Agriculture is one of the biggest employers of women – 65% of all women who work, work in agriculture [115]. However, despite the disproportionate amount of physical labor performed by Pakistani women, they are often locked out of decision-making at all levels [10]. In major water-related bureaucracies, women make up only 2-5% of the staff and occupied 0 senior positions [10]. And households are traditionally patriarchal, where men make most decisions regarding the family.

In rural high-mountain communities, the responsibility of managing household water resources often traditionally falls on women [10]. Because of the difficult terrain, infrastructure in the mountains is under-developed compared to the lowlands; therefore, in many areas in the H-K, it is customary for women to travel long distances to fetch water and fodder for fuel for their households [158]. In times of food scarcity, they are often given less to eat and drink, leading to an increased tendency for malnutrition and health problems [60]. This exacerbates the physical labor that they must perform, putting additional strain on their bodies, which already carry and give birth to children [60]. A natural consequence of increasing women's workload is withdrawing children from school to assist with tasks; female children are usually the first to be withdrawn from school [60]. Preventing girls' access to education is well-known to cause additional gender inequities because lack of education disallows women to pursue more high-paying jobs and achieve economic independence.

2.3.3 Social Vulnerabilities and Strengths

Climate disasters often increase women's vulnerabilities. Women and children are 14 times more likely to die during climate-related disasters [115]. When water is scarce, women face otherwise avoidable health issues such as urinary tract infections and uterus prolapse from carrying heavy loads of water over long distances [10]. When disasters hit, they are also put in charge of taking care of children and livestock, and are responsible for the greater part of emergency preparedness, leading them to neglect their own needs [14].

Despite this, gender equality is slowly improving both in highland communities and in South Asia in general. A generational divide in the perceptions of women's strengths and weaknesses in Hunza Valley, Gilgit-Baltistan demonstrates how education is improving women's confidence and sense of agency [83]. In this community, which is highly vulnerable to GLOFs from the nearby Shishper Glacier, both men and older women said in interviews that menstruation and pregnancy made women physically weak [83]. Older women also said they were not emotionally or physically prepared for disaster because they were too weak to run and be agile. Their confidence was limited because of their financial dependence on the men in their households and lack of access to emergency resources [83]. One woman said, "I think women cannot do anything unless men are...there for them. They cannot make any decision on their own; they are not as brain smart as men" [83]. However, a stark difference in attitude emerged in young women aged 17-25; many were college-educated and believed "We are equal to boys...We are educated. We can understand any situation, make decisions, and act accordingly without much fear" [83]. One man even said, "Women nowadays are more competent than men. If they are surpassing men in different fields, including disaster management, there is nothing wrong with it. If anything, we appreciate and encourage our daughters" [83]. Older women also acknowledged the role that education had played in the empowerment of their daughters and granddaughters, and expressed that if they had had access to trainings and education, they may feel less paralyzed and afraid during disasters [83].

Women are usually the main disseminators of information amongst other women [115]. They are more likely to take emergency precautions than men and carry important ancestral knowledge that help them prepare for emergencies [83]. When women are empowered during disasters, it allows men to be less stressed and over-stretched during disasters [83]. The importance of education for girls and women cannot be overstated; it is the foundation for reaching gender equality and better community resilience to environmental disasters. It is therefore essential for women's needs and participation to be prioritized by projects like GLOF-II that provide trainings and awareness sessions to communities affected by GLOFs.

2.4 Chapter Conclusion

Pakistan only contributes 1% of anthropogenic carbon dioxide emissions, but it is the eighth most vulnerable country to climate change [121]. The fifth most populous country, it is home to the largest number of glaciers of any country on Earth. However, it is suffering greatly from the effects of climate change.

In Pakistan, catastrophic floods from glacial lakes and monsoon rains have caused billions of dollars in damages and indescribable pain and suffering. In parallel, the growing problem of water scarcity looms over Pakistan's agricultural industry. Due to climate change, it is expected that current systems in place to deal with water insecurity will not be sustainable in the long term. Scholars have posited that watershed-level approaches are necessary to manage the Indus basin, which feeds most of Pakistan. However, because the Indus is shared by multiple countries, governments will have to work together to share resources multilaterally rather than operating under the assumption that each of their interests are competing [137]. Due to the political tensions between nations whose borders fall within the H-K region, the creation of such types of agreements is both unprecedented and may be unlikely to occur in the near future.

The hydrology of the H-K region is of great importance to the water systems of Asia because most of the continent's major river systems originate from these mountains. As such, the stewardship of water resources in the mountain highlands is extremely important for the health of the water systems downstream in the lowland plains. The stress of water and consequent food shortages mainly falls on women due to patriarchal family structures and gender roles; climate-related disasters also make women even more vulnerable. In particular, increasingly severe GLOFs in the northern regions of Pakistan have caused widespread destruction, displacement, and mental health crises. The education of girls and women is critical for better disaster response and for their own safety, wellbeing, and mental health.

Multilateral, inter-national cooperation is the only way to ensure that all people who rely on the water systems originating from the Himalayas will have access to water in the future. As of today, mutually beneficial multilateral agreements between governments in South Asia have been rare; however, because water is a shared resource, cooperation is in the self-interest of every South Asian nation.

Chapter 3

Glacier-Related Hazards

In this chapter, we will discuss two types of glacier-related hazards that have been commonly observed in the Himalayas: surges and GLOFs (glacial lake outburst floods). While GLOFs generally occur over short timescales (i.e., hours to days), surge behavior can occur over months to years. GLOFs are therefore much more destructive to human lives, and are notoriously difficult to predict. Although surges pose less of an urgent danger to downstream residents than GLOFs, they are intricately linked with GLOFs due to their ability to create new glacial lakes and change the underlying landscape. Understanding surge behavior is therefore also important in predicting GLOFs and how glaciers may respond to climate change.

3.1 Anatomy of Mountain Glaciers

Alpine glaciers typically exhibit a general "anatomy" which is primarily composed of two zones: the accumulation and ablation zone (Figure 3-1). The accumulation zone is generally at a higher elevation than the ablation zone; the colder temperatures at higher elevations reduce melting in this zone. In contrast, warmer temperatures at lower elevations mean that greater mass loss occurs in the lower-elevation ablation zones [125]. The line between the accumulation and ablation zones is called the equilibrium line or the firn line [125]. Between winter and summer seasons, the equilibrium line may move up and down as the size of the accumulation and ablation zones change.

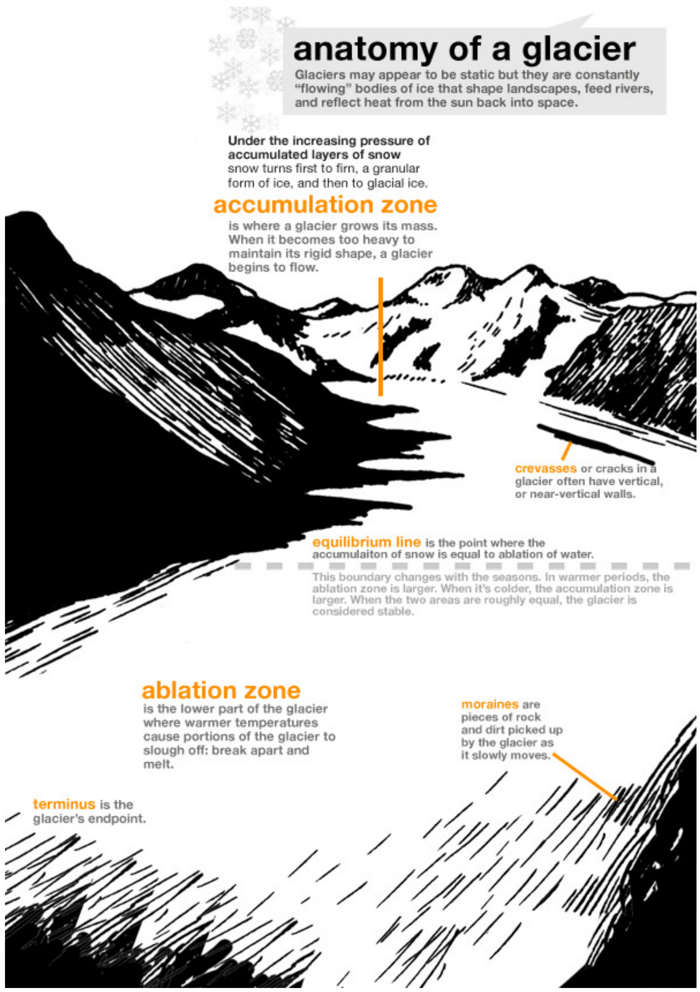

Figure 3-1: *Anatomy of a glacier [125]*

For the Karakoram region in particular, movement of snow via avalanches due to steep and rugged terrain can complicate the definitions of mass accumulation; this is sometimes referred to as "avalanche nourishment" [65]. These glaciers tend to start with steep icefalls and avalanche cones which converge to the main ice mass of the glacier [63]. Glaciers that are avalanche-nourished often have no clearly delineated accumulation zone [65]. The extremely rugged, avalanche-prone terrain of Karakoram glaciers, as well as their remote location and high elevation, makes establishing in-situ monitoring programs very difficult. This leads to a major scarcity of data that makes studying the region challenging.

3.2 Surges

Glacial surges are one of glaciology's greatest mysteries. Surging is a rare but significant phenomenon that affects about 1% of glaciers worldwide [140]. During surges, glaciers move at uncharacteristically high speeds 5-100 times their normal speed, often appearing to be unrelated to external triggers [96]. These periods of movements also cause significant advancements of the terminus of the glacier. Surges are separated by longer, "quiescent" periods and often occur cyclically with semi-regular intervals between surging events [96]. It is generally agreed that surges occur due to some sort of hydrological, thermal, or mechanical conditions or properties at the bed of the glacier; this is supported by the fact that surges have been observed to cluster in geographic areas [96]. Clusters of glaciers that exhibit surge behavior fit in well-defined "climatic envelopes" that exhibit optimal temperature and precipitation ranges [140]. Glacier size and slope may also play a role in surging, with longer and larger glaciers as well as shallower slopes showing significant correlation with surging activity [140]. Areas where glacial surges have been commonly observed include but are not limited to Alaska-Yukon, Arctic Canada, parts of Greenland, Iceland, Svalbard, and in mountain ranges such as the Tien Shan and Karakoram in Asia [140]. However, the wide diversity of surge-type glaciers suggests that many factors can contribute to surge behavior [140]. Understanding the changes in basal conditions that lead to surging is a difficult task, but an increased understanding of surging has the potential to elucidate a variety of subglacial processes that affect estimates of glacial runoff and sea level rise.

3.2.1 Suggested Mechanisms for Surging Behavior

Most existing studies that attempt to explain the phenomenon of surging have focused primarily on the hydrological and thermal aspects of basal conditions. Surging can be described as a cyclical process during which a force resisting the glacier flow allows enough glacial mass accumulation to initiate the next surge, during which the resistive energy is released [51]. The resistive processes that dominate the quiescent periods between surges, as well as the factors that trigger surges, could be hydrological (for example, [80]) or thermal – or, more likely, some combination of the two. Generally, the time between surge cycles depends on the rate of accumulation of mass between surge events; a critical potential energy must accumulate in order to trigger another surge [92].

Thermal Conditions and Enthalpic Balance

Studies suggesting a thermal trigger for surging usually propose that rises in temperatures at the base of ice sheets trigger surging motions [126]. Strain heating, which increases as mass accumulation increases the overburden pressure of the glacier, can cause instability and surging when the basal ice is close enough to the melting point [38]. Polythermal conditions at the bed may indicate instability, and the transition zone between cold and warm regions may correspond to the surge front [105]. After basal melting is initiated, water penetration from the subglacial till into crevasses warms the overlying ice, causing further instability [37]. These hypotheses maintain that surges are stopped by periods during which the bed is frozen to the ice. However, clusters of surges have been observed in glaciers in both temperate and colder areas. Surges have also been observed to stop at the peak of summer, when it is clear that the ice is not frozen to the bed (e.g., [26]). Therefore, these proposed thermal triggers require further investigation and are very unlikely to be the sole factor in the initiation of surges.

More recently, enthalpy balance equations have been used to explain surge mechanisms [140]. The idea of an enthalpy-based model encapsulates both latent and sensible heat and potential energy [18]. In glaciers, total enthalpy represents the internal energy of the glacial system, which is a function of ice temperature, volume, and water content [18]. As glaciers flow downwards, gravitational potential energy is converted into frictional heat, which can be in the form of either latent heat (i.e., heat released or stored in a phase change) or sensible heat (i.e., heating that does not result in a phase change) [140]. This process triggers basal melting and increased lubrication at the bed, causing higher velocities in a positive feedback loop [18]. Mass balance equations can also lead to the calculation of a "balance velocity," above which dynamical thinning will occur and below which the glacier will thicken.

[140] also suggest that surges occur in "climate envelopes", or specific ranges of temperature and precipitation, due to mass and enthalpy balances straying from steady-state conditions. Since heat conduction to the atmosphere and subglacial drainage are not efficient within the "optimal surge envelope" of temperature and precipitation, the glacial system oscillates between slow and fast movement to dissipate excess energy [140]. Although these enthalpy and mass balance models are fairly simplistic, they provide a good foundation for future modeling of conditions leading to surges.

Mechanical Conditions

Despite the prevalence of subglacial till underneath surge-type glaciers, few studies have been done on the mechanical drivers of surge instabilities. Minchew and Meyer provide an insight into the mechanics of incipient glacier motion without taking meltwater flux or enthalpic factors (which, when taken alone, cannot fully explain surging) into account [100]. The process starts with a process called dilation, during which grains of till sediment roll over each other, moving from a close-packed arrangement to one that has much more pore space. The dilation of the granular, porous subglacial till decreases the pore water pressure, creating a suction effect which strengthens till and prevents surges [100]. Incipient motion occurs when dynamical thinning of the overlying glacier happens faster than the drops in pore water pressure: dynamical ice thinning decreases the overburden pressure, which reduces effective stress (i.e., the contact stress between grains of till) [100]. Since the change in pore water pressure is happening slowly, it is not able to compensate for the decrease in ice overburden stress, which is what allows the effective stress to decrease. The decrease in effective stress decreases the shear strength of the till, which results in initiation of surging. This mechanism depends on the till having low enough hydraulic permittivity to allow drops in pore water pressure to occur slower than dynamical thinning. Because this study does not take into account the evolution of the subglacial hydrological system (i.e., channels and drainage systems that may form and close), it cannot explain the termination of surges, but provides a good insight into mechanical processes that may explain incipient surge motion.

Subglacial Hydrology

Subglacial hydrology has provided more compelling explanations for surge behavior. The prevailing theory maintains that the transition between an efficient, channelized system to an inefficient "linked cavity" system drives up water pressure and lubricates the ice-bed interface, triggering a surge [80]. During surge events, a distributed drainage system maintains a high basal water pressure, while quiescent periods are characterized by efficient, rapid transportation of subglacial water [27]. The downside of this explanation is that it is somewhat simplistic and does not account for how incipient surge motion is initiated or terminated. A further discussion of the role of subglacial hydrology in surging will be presented in Chapter 4.

3.2.2 Surges in the Karakoram

The Karakoram region of the Himalayas is one of many areas where clusters of surges have been observed [140]. The glaciers of the Karakoram region in particular are an extremely important source of water for local ecosystems and people [29]. It is crucial to form a better understanding of glacial surges in the Karakoram because they have strong implications for landscape evolution [70] and frequently reroute meltwater and create ice-dammed lakes which can lead to dangerous glacial lake outburst floods (GLOFs) [22]. The interactions of surges and the dangers of GLOFs will be discussed next in Section 3.3.

Studies of the Karakoram have been conducted by Western scientists since the 1800s (e.g., [58]). In the past century, surges in the Karakoram have been observed to destroy villages, forests, and agricultural land, causing terror to local residents [44]. Although the spatial occurrence of surges has remained relatively constant, the temporal frequency has increased since the 1960s, likely corresponding with the

Karakoram's slight positive trend in overall mass balance [41]. As climate change progresses, it is likely that surge behavior will become more common amongst glaciers [41]. Indeed, a recent surge of Shishper Glacier in northern Pakistan, combined with an unusual warm spell, created conditions that led to a massive GLOF that destroyed a major highway in Gilgit-Baltistan, Pakistan [26].

Although some have suggested that surge-type behavior is primarily exhibited by tributary-type glaciers (e.g., [64]), other studies have disagreed as to whether the Karakoram's surges are largely thermally or hydrologically controlled [120]. It is also suggested that a continuum of basal conditions exist across the Karakoram [64]. As a result, it is likely that a combination of environmental factors triggers surge activity; short-term instances of high-altitude warming likely trigger clusters of surges [64] which seems plausible given that occurrence in certain "climatic envelopes" has been shown to correspond strongly with surge activity [140].

3.3 Glacial Lake Outburst Floods

Glacial lake outburst floods (GLOFs) are outburst floods caused by failure of dams that contain glacial lakes. GLOFs are dangerous phenomena during which millions of cubic meters of water rush downstream, destroying infrastructure, livelihoods, and communities below [94]. Examples of the suffering that have been caused by GLOFs have been discussed in Chapter 2.2.

GLOFs have been observed to occur globally wherever there are alpine (mountainous) glaciers. The prevalence of glacial lakes and the occurrence of GLOFs have both increased greatly in recent decades [94], as warming temperatures contribute to greater glacial melt and larger glacial lake volume. However, GLOFs have proven difficult to study due to their unpredictability and the difficulty of making in-situ observations in remote and dangerous environments. Therefore, much of the study of GLOFs today relies on remote sensing and modeling.

GLOF inventories have provided a comprehensive way to keep track of changes in glacial lake frequency, size, and volume - for example, [113] and [94]. These inventories are created using satellite imagery via methods such as supervised and unsupervised image classification or thresholding of indexes such as NDWI which identify bodies of water [68]. Some studies have also used the shallow-ice approximation (SIA), which only considers gravitational driving stresses and basal drag, to predict overdeepening sites where glacial lakes may form [144]. Other studies predict the potential impact of GLOFs by applying multi-factor risk criteria to each glacial lake [144, 133, 148]. Finally, hydrodynamic modeling using engineering softwares such as HEC-RAS can be used to investigate flow paths and predict impacts of glacial lakes should they undergo dam failure [133].

In particular, the retreat of glaciers in the central and eastern Himalayas has led to greater accumulation of meltwater between the glacier terminus and the frontal moraine of many glaciers [94]. The entrainment of sediment as the flood progresses downstream can increase the flood volume by orders of magnitude [134]. Although the greatest number of GLOFs have been observed in the eastern Himalaya, they also occur frequently in the Karakoram region and are frequently associated with surging there (e.g., [128, 26]). Furthermore, GLOFs are predicted to increase in frequency and severity in the Karakoram [22]. This is a region that is already vulnerable to damage due to high population densities near the mountains. Therefore, it is of the utmost importance to improve our understanding of GLOFs both in the eastern and Karakoram regions of the Himalayas so we can predict the timing and severity of these events.

3.3.1 Classification of Glacial Lakes

Glacial lakes can be dammed by a variety of materials, including glacial moraine of varying strengths (i.e., “moraine-dammed”) or other glaciers (i.e, “ice-dammed”). In both cases, meltwater from the source glacier is forced to accumulate behind the natural dam, forming a glacial lake that can be more or less stable depending on the mechanical strength of the dam and the pressure exerted on it by the lake water [111]. While other types of glacial lakes exist, such as kettle lakes, we will neglect them in our summary here because they exist in valleys rather than in the mountains and do not exhibit outburst flooding.

Ice-dammed lakes

Ice-dammed lakes are a common category of glacial lakes. Aptly named, they are dammed by ice [111]. In this work, when we refer to "ice-dammed lakes" we speak of lakes that are dammed by another glacier.

Jökulhaups are a particular type of GLOF resulting from an ice-dammed subglacial lake, generally occurring in Iceland, with the most famous ones occurring in the Vatnajökull ice cap [28]. In general, the volumetric discharge of jökulhlaups increases slowly over time, until it peaks and then abruptly stops. Jökulhlaups usually originate from marginal, englacial, and subglacial lakes, meaning lakes that are contained inside or underneath a glacier [28]. Generally they are caused by permanent geothermal heat sources or volcanic activity [28].

Moraine-dammed lakes

A great number of glacial lakes worldwide are moraine-dammed [111]. Glacial moraine, or till, is the highly heterogeneous rocky and sedimentary material left behind from glacial movement [129]. Due to its heterogeneity, the strength of moraine dams is difficult to measure and predict [36]. Mechanical failure can result from sedimentary erosion, also referred to as "piping", or by overtopping of the dam [24].

The most common mechanism leading to glacial lake outbursts is by a breach of the moraine that dams the lake [24]. The occurrence of a dam breach is dependent on both the stability of the dam and the probability of a triggering event [132]. The moraine dams can fail by two common mechanisms: 1) overtopping, and 2) piping due to seepage [24]. Failure generally depends on the frictional qualities of the moraine and the pressure of the lake water causing internal erosion.

Most studies on moraine dam failure focus on overtopping rather than seepage failure. GLOFs resulting from moraine dam failure are usually triggered by events such as avalanches, calving from the glacial snout, earthquakes, and severe precipitation [111]. These events result in rapid volume transfer into the glacial lake that results in a series of seiche waves that can lead to overtopping if the dam freeboard (i.e., vertical distance between the lake surface and crest of the dam) is small [157]. Other volume-transfer events include influx of water from storms as well as the release of subglacial or englacial reservoirs [157]. In addition, earthquakes can cause mechanical failure of moraine dams which lead to catastrophic flooding downstream. However, [148] note that due to climate change-driven destabilization and melting of permafrost and steep glaciers, avalanches will occur more frequently in the future and must be closely studied as they will be a dominant trigger for glacial outburst flooding in the future. Climate change may also drive the degradation and melting of ice-core moraines, which will weaken the moraine and cause the crest of the dam to sink, thus decreasing the freeboard and increasing the risk of flooding [157].

3.3.2 GLOFs in the Karakoram Region

A particular type of GLOF that is largely unique to the Karakoram, as opposed to the eastern Himalayas in India, Nepal, and Bhutan, originate from ice-dammed lakes associated with advancing glaciers [66]. Mostly occurring in the upper Indus and Yarkand basins, these ice-dammed GLOFs have a great potential for destruction [66]. Although a spectrum of ice-dammed, moraine-dammed, and mixed ice and moraine-dammed lakes exists in the Karakoram, all of the largest most dangerous floods reported have originated from ice-dammed lakes [66]. Most of these are comprised of a main lake dammed by ice from a "lateral valley" [66, 71]. Many glacial lakes in the Himalayas of India, Nepal, and Bhutan are caused by retreating glaciers; however, these Karakoram lakes are caused by impoundments of meltwater by advancing glaciers. Because of this, these lakes are often short-lived and only last for months or years, as opposed to other glacial lakes which accumulate in static depressions in the landscape or ice [66]. They are more often characterized by sudden drainage under the ice dam rather than overtopping or drainage around the margins of the dam [66]. In the upper Indus basin, no floods have been observed as a result of sudden mass transfer into the lake – i.e., avalanches or landslides – suggesting that factors internal to the glacier system are at play [66]. For example, it has been suggested that water pressure exceeding the yield strength of the ice dam may cause subglacial drainage, but there is little evidence to support this claim [57]. It is more likely that these lakes drain through exploitation of already existing drainage systems [66]. In fact, it is hypothesized that the drainage of water through existing subglacial hydrological systems leads to more rapid drainage and larger outburst floods [66].

The prevalence of glacial surge behavior in the Karakoram leads to an increased incidence of GLOFs and smaller drainage events from temporary ice-dammed lakes. These GLOFs and drainage events can have devastating effects on the communities downstream (discussed in Chapter 2).

3.4 Chapter Conclusion

GLOFs and surges are connected phenomena that occur commonly in the Karakoram range of the Himalayas. Many GLOFs in the Karakoram come from ice-dammed lakes that are created through the impoundment of meltwater by a surging glacier in a lateral (adjoining) valley. Surges are one of the most mysterious phenomena in glaciology, and are poorly understood. There have been a variety of theories as to how they start and stop, which we have discussed in Section 3.2.1. While none of the proposed mechanisms can fully explain surge behavior, it is likely that it is caused by a combination of mechanisms and environmental conditions.

Although they occur commonly in moraine-dammed lakes in the eastern Himalayas, most GLOFs in the Karakoram come from ice-dammed lakes. Even more so than moraine-dammed GLOFs, it is likely that ice-dammed GLOFs occur due to some subglacial hydrological factors. However, very little is known about what really happens when these lakes burst through their ice dams. The field of subglacial hydrology may be able to shed more light on the factors behind such rapid lake drainages.

Chapter 4

A Primer on Subglacial Hydrology

Subglacial hydrology is the study of the flow of water beneath and through a glacier. A growing field within glaciology, subglacial hydrology can provide insights into the movement of glaciers and the occurrence of glacial hazards. In particular, subglacial hydrology provides an important control on sliding velocity of glaciers. This is highly important in the context of sea level rise; subglacial hydrology has been shown to play an important role in how fast Greenlandic and Antarctic ice sheets slide into the ocean, thus accelerating rising oceans [162]. In addition, changes in subglacial hydrology are implicated in surging behavior [128, 25, 100] and glacial lake outburst floods (GLOFs) [128].

Subglacial processes are still very poorly understood by science, especially in alpine environments. Most existing observations of sub- and englacial hydrology are based on studies in Greenland and Antarctica and must be extrapolated to alpine environments due to the difficulty of obtaining in-situ measurements, especially in High Mountain Asia [150]. As a result, models must attempt to fill our gap in understanding in the alpine glacial environment until more observations are available. In this chapter, we will discuss methods for observing and modeling subglacial hydrology, as well as our current understanding of the physical systems themselves.

4.1 Subglacial Drainage Networks

The seasonal evolution of the subglacial drainage network is a fascinating process with many feedback loops. Much of the evolution of these systems is driven by changes in temperatures across seasons. Understanding how subglacial hydrology modulates the seasonal speed-ups of glaciers can help us understand and predict ice mass balance in the long term [124]. In addition, it can help understand the seasonal timing of GLOFs and surges. The processes governing the evolution of subglacial drainage networks are discussed below.

4.1.1 Delivery of Meltwater to the Bed

During summer melt seasons, a great amount of melt is generated at the surface of glaciers [21, 162]. In 1986, Iken and Bindschadler showed that there is a strong and instantaneous coupling of surface and basal meltwater in alpine glaciers [72]. [141] used GPS tracking and inSAR remote sensing data to show that the transportation from the surface to the bed is fast enough that it can cause basal lubrication even across the thicker Greenland glaciers. [141] also showed that supra-glacial lakes may play a crucial role in "priming" the conduits that link the surface and base of ice sheets.

The strong coupling between surface and basal meltwater environments has provided justification to make model assumptions that all meltwater input to the bed (i.e., surface melt, rainwater, aquifer contributions) are instantaneous. Indeed, this assumption is employed in the SHAKTI model, which will be described in Section 4.3.1. However, englacial water storage can delay the delivery of surface water to the base. Observations in Greenland have revealed that long-term water storage happens in englacial firn aquifers, especially in glacial accumulation zones [82]. In addition, water can remain at the surface in damaged crevasses as observed in Store Glacier in Greenland [82]. This water can remain for months to years [34], and may or may not be hydraulically connected to the subglacial drainage system [82]. Observational studies of englacial water storage have seldom been done in alpine contexts, and never before with radar sounding. Because of this, we do not know much about the size or capacity of firn aquifers in alpine glaciers.

Seasonal meltwater storage can also occur in moulins (supraglacial lakes) on the surface of glaciers [42]. In Greenland and Antarctica, water pressure exerted by moulins has also been observed to cause full-thickness hydrofractures that allow delivery of the full or partial lake volume to the subglacial environment [35]. The nature of these hydrofractures is connected to the stress and velocity regime of the glacier, as well as the presence of pre-existing moulins [35]. In addition, a model developed by [85] suggests that lake hydrofractures dominate surface drainage pathways at higher elevations in West Greenland, but crevasses drain most of the surface water at lower elevations. Regardless of the complexities of hydrofracture formation in ice sheets and glaciers, it is clear that they provide a critical hydraulic link between the surface and the bed.

It is very likely that some long-term water storage occurs in Himalayan glaciers, as the presence of moulins and damaged crevassed areas have been well-documented there. However, there have been few to no studies on the degree of supraglacial and englacial water storage in alpine environments, so the timescale and volume of this storage is largely unknown.

4.1.2 Components of the Subglacial Drainage Network

The processes of channel formation and drainage network evolution are of great relevance in this work, as we model the evolution of the subglacial network in Chapter 5. The opening and closing of channels are governed by material properties of the ice and bed, while the nature of the drainage system is controlled by complex feedback loops governed by spatial and temporal water pressure gradients. The various modes of subglacial transport (channels, canals, sheets, cavities, and porous flow) are shown in Figure 4-1.

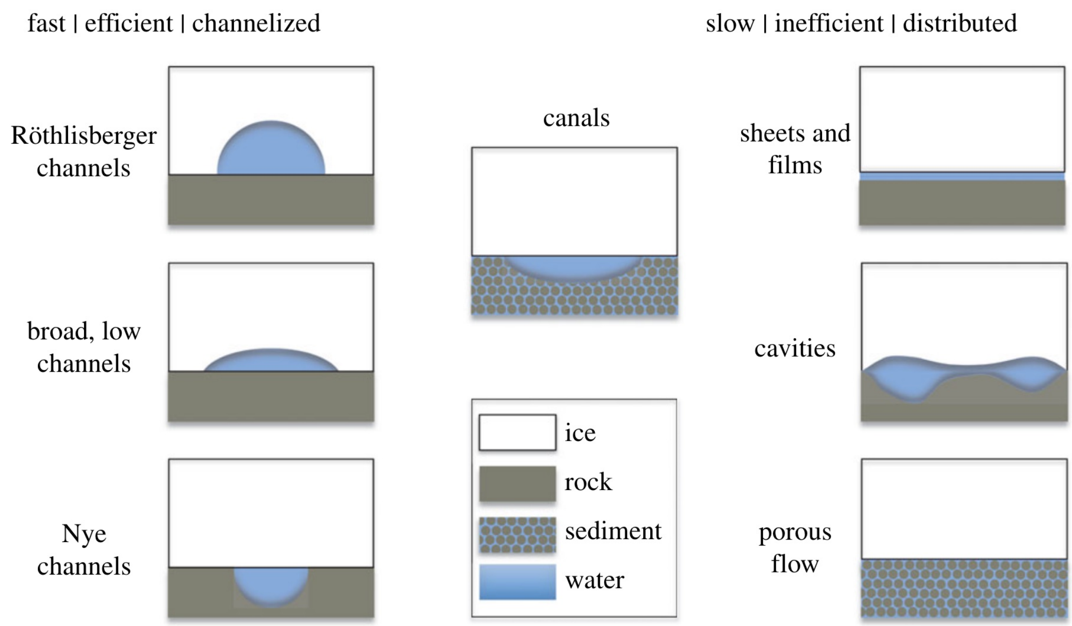

Figure 4-1: *Network types as illustrated by [50]. Efficient drainage networks are characterized by clear-cut channels, while inefficient drainage networks are characterized by sheetlike or distributed drainage geometries. Canals, which contain both channels and porous media, exist in a gray area between the two classifications [50]. Efficient vs. inefficient systems are discussed in 4.3.*

Röthlisberger and Nye Channels

Röthlisberger first introduced the idea of subglacial water channels cut into the ice in 1972 [130]. These channels are often referred to as R-channels. This idea was then elaborated upon by Nye, who proposed channels incised in bedrock (N-channels) [114]. Röthlisberger channels, which are either circular or semicircular, grow or shrink based on the balance between 1) opening of the channel by melt and 2) the closing of the channel by inward creep [52]. This creep is governed by Glen's flow law $\dot{\epsilon} = A\tau^n$, a constitutive relationship describing the strain rate of ice based on its material properties and stress regime. Melt opening of the channel is governed by frictional heating from turbulent water flow [130]. The turbulent water releases gravitational potential energy as heat, and the resultant melt opens the channel. The consequent melting is dependent on water flux, effective pressure, and the latent heat of fusion for water.

Cavity Formation and the Linked-Cavity System

Cavities can form when ice flows over rough bedrock. The spaces are created on the leeward (downstream) side of any small obstacles, and can become part of the subglacial drainage network when they are filled with water [52]. The existence of cavities in a subglacial system can allow for a linked-cavity system rather than a channelized drainage system which is solely made of Röthlisberger channels [80]. The linked-cavity system, as the name suggests, consists of large cavities linked by small drainage channels between them [52]. The main difference between a channelized vs. linked-cavity system is the relationship between effective pressure N and water flux Q: for Röthlisberger channels, $\frac{\partial N}{\partial Q} > 0$, but for the linked-cavity system, $\frac{\partial N}{\partial Q} < 0$. Let's take a closer look at why this relationship is different for each system.

For channelized systems (i.e., Röthlisberger channels), increases in flux lead to channel opening, since the melt rate m is proportional to the flux:

$$mL = Q\left(\phi + \frac{\partial N}{\partial s}\right) \tag{4.1}$$

where L is the latent heat of fusion and N is the effective pressure. As the channel widens with increasing flux, the water pressure p_w decreases since there is now more space in the channel. Because lower-pressure water does not support as much of the overburden weight, there is an increase in the effective pressure N. Therefore, we see a positive relationship between flux Q and effective pressure N, and a negative relationship between channel water pressure p_w and flux Q. As a result, Röthlisberger channels tend to form in arborescent (tree branch-like) networks, where larger, low-pressure channels form preferentially and can out-compete smaller, high-pressure ones. This can be seen clearly in valley glaciers, where steeper pressure gradients at the lateral margins give rise to a system of a few very large channels fed by smaller tributaries [50].

In contrast, the linked-cavity system exhibits a negative relationship between effective pressure and channel water flux. The cavities store most of the water in this type of system, and channels between them play a smaller role. It is then a little more simple to see that when the flux through a cavity increases, cavity pressure also increases, thus reducing the overburden (effective) pressure N.

The channelized system and the linked-cavity system are two idealized end-member cases - in real life, we can see a variety of systems that fall in between the two cases. Since water flows down effective pressure gradients (i.e., high to low water pressure), the system will be in equilibrium if the effective pressure is the same everywhere. However, the system can switch or move between states depending on $\frac{\partial N}{\partial Q}$. An important factor to consider is that the effective pressure N for the linked-cavity system is dependent on sliding velocity u - so if u is large enough, the system will move from channel drainage to cavity drainage [54]. For a more detailed mathematical explanation of these processes, please reference [52], pages 55-61.

Soft and Hard Beds

Until this point, we have primarily considered subglacial hydrology as understood for a rigid underlying bed. This is because most studies (both observational and modeling) of seasonal evolution of subglacial hydrology consider only the case of a rigid bed. Assuming that a glacier bed is rigid (i.e., solid bedrock) means that all subglacial drainage channels are cut solely into the ice rather than the ground below. However, many glacier beds are actually made up of deformable till and sediments, which also can hold and store water and be eroded to form channels [17]. This is because glaciers tend to cause erosion of the bed: they generate boulders which are ground into smaller rocks and a range of sediments of various sizes [52]. This heterogeneous mix of rock and sediment is known as glacial till.

The presence and properties of deformable till have many implications for glacier and ice sheet flow and instability (for example, in surge behavior described in Section 3.2.1). The sliding velocity of the ice can strongly effect the subglacial hydrology - for example, faster velocities can lead to cavity formation. However, the properties of the till itself have important implications for subglacial channelization in a soft bed environment.

It has been observed that creep of till and transport of sediment plays an important role in controlling sediment channelization beneath glaciers [16]. Unconsolidated sediment is widely observed under glaciers in both Greenland and Antarctica (for example, [135, 86, 52]). The subglacial network and its seasonal evolution is heavily dependent on the properties of the underlying till, as glaciers resting on soft beds can have canals, macroporous films, and anastomosing channels [61]. This generates a spectrum of complicated relationships between basal sliding, till water pressure, and the resultant properties of the subglacial drainage network [61].

Whether or not the till can deform is dependent on the Coulomb yield criterion, which describes failure at yield stress τ_0 of an isotropic material with cohesion c_0, effective pressure N, and friction angle ψ:

$$\tau_0 = c_0 + N \tan \psi \tag{4.2}$$

If this critical yield stress τ_0 is reached, the material will deform. When N increases, the frictional resistance to flow will increase, thus causing the strain rate $\dot{\epsilon}$ to decrease. [31] proposes a power law relation for this behavior:

$$\dot{\epsilon} = A(\tau - \tau_0)^a N^{-b} \tag{4.3}$$

where a and b are positive, non-dimensional parameters and A is a constant dependent on pressure. Here we can also see that strain rate $\dot{\epsilon}$ increases when N decreases, and vice versa. More shear stress also causes a higher strain rate in the till.

To conclude, when the yield stress of till is exceeded, the material undergoes plastic deformation, which can result in canal or channel formation. This yield stress is dependent on the properties of the till as well as effective pressure N. Of course, N, which describes the contact force between till particles, is also dependent of the pressure of the water in the till.

Sedimentary Canals

Canals are channels that have an ice roof (i.e., they are underneath the glacier) and a sediment floor (Figure 4-2). When modeling this system, we consider both

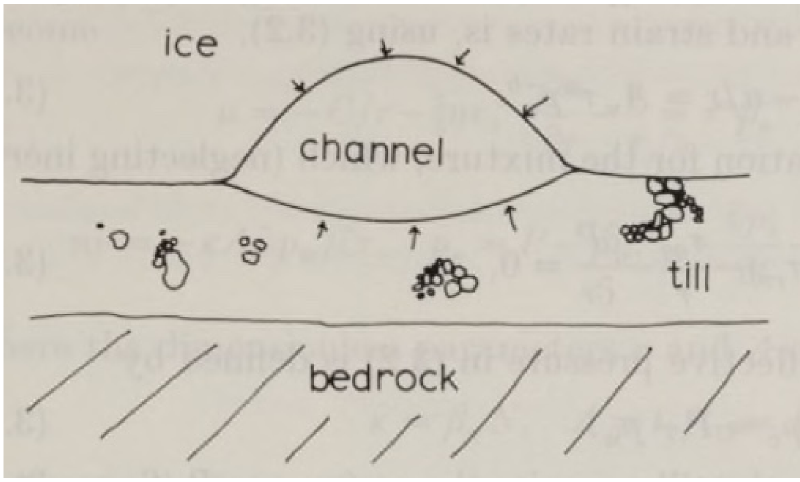

Figure 4-2: *Cross-sectional illustration of a canal (ice roof, till floor) from [53]*

melt opening and ice creep on the roof, and sediment erosion and till creep on the floor of the canal [153]. Mathematical analysis of these processes suggests that for soft-bedded valley glaciers, Röthlisberger channels may dominate the steeper areas, while shallower slopes will favor canal drainage [50].

Besides sediment creep and transport, which are complicated processes on their own, porous flow must also be considered. The flow of water in a porous medium goes from areas of high hydraulic head (potential) to low hydraulic head. However, this is limited by the transmissivity K of the material, so that flow is governed by

$$\frac{\partial h}{\partial t} = \frac{K}{S_s}\nabla^2 h \tag{4.4}$$

where h is hydraulic head, t is time, and S_s is "specific storage", a parameter dependent on aquifer porosity and compressibilities of the aquifer and the fluid [50]. We can see from Equation 4.4 that water will flow faster where the hydraulic transmissivity is higher and the difference in head is larger.

All of this is presented to the reader simply to demonstrate that the presence of a soft deformable till underneath a glacier can introduce non-trivial effects into the subglacial environment. We will now see how this can affect the evolution of subglacial drainage from season to season.

4.1.3 Seasonal Evolution of the Drainage Network

Numerous studies have shown that the velocity of glaciers increases during melt seasons (for example, [124, 162, 61, 26]). It has been well-established that these summer speed-ups are due to an increased delivery of water to the bed of the glacier, which increases water pressure, lubricating the ice and the bed [162, 62]. However, bed lubrication has a more complicated relationship with the amount of meltwater reaching the bed, since the subglacial network adapts and responds to changes in meltwater input.

During the winter, drainage paths are closed because there is very little water flow to resist creep closure [62]. This gives rise to a system with low hydraulic transmissivity with few channels and connections. Spring and summer bring warmer temperatures, which lead to surface melt that migrates down to the bed [62]. The melt at the bed is very high-pressure due to the low transmissivity of the drainage network – the water has nowhere to go. This high-pressure environment leads to bed lubrication and speed-ups as mentioned earlier (the mechanisms that link high-pressure lubrication to higher sliding velocities will be elaborated upon in Section 4.1.4). [61] has also proposed that basal water is stored in the subglacial environment during the melt season; as the water pressure increased, they observed that discharge from the bed increased as the system reorganized. They also observed that water was stored in subglacial sediments, cavities, and the braided system of channels during the summer - see Figure 4-3 . Diurnal (day-scale) variations have also been thought to be accommodated by moulins [42]. Along with fast flow comes shear-induced till dilation and glacial uplift [61].

The late summer is quite different from the spring and early summer. Recalling Equation 4.1, we understand that melt opening is increased with greater flux: as more water enters the system, more channels are opened and the system becomes more efficiently connected. So long as high fluxes continue, melt opening will exceed creep closure, thus keeping the channels open. This means that the drainage system is now able to quickly shuttle large fluxes through, which allows it to return to a low-

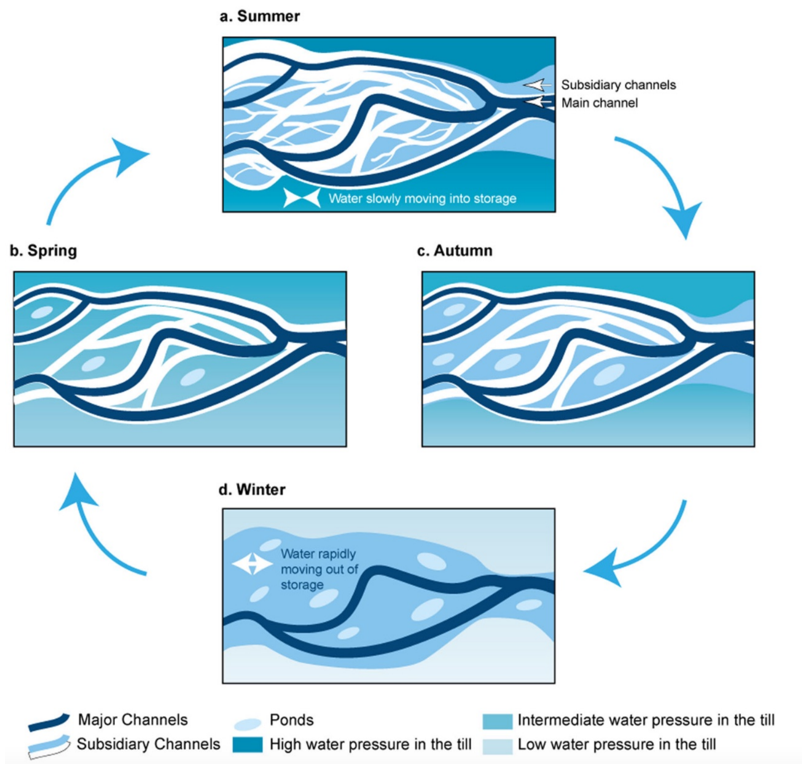

Figure 4-3: *Schematic diagram of the seasonal evolution of channels in a soft-bedded glacier in Iceland, presented by [61].*

pressure state. This then causes a reduction in lubrication and slow-down of sliding velocities. Every now and then, when the input exceeds the increased capacity of the summer system, it may return to a high-pressure state [61]. In the summer, when high shearing occurs, the till undergoes cycles of dilation and compaction. During dilation, pore spaces grow and the pressure inside them decreases, pulling in additional water into the till [100]. When the glacier stops moving and the till undergoes compaction again, this water is discharged from the till [61].

As temperatures become colder in the autumn to winter seasons, surface melting decreases and so does the flux through the subglacial system. Creep closure becomes the dominant mechanism and the system transitions back to the less-connected state. As the system contracts, water is discharged from the ice-bed interface [34]. Because it does not have the capacity to handle much flux, the winter state is more sensitive to short (dayslong) periods of warmth during autumn and winter, which can lead to abrupt drainages [61].

4.1.4 Sliding Laws

Basal slip is an important boundary condition on ice sheet models; unlike for ice shelves, which float on water, ice sheets and glaciers lie on solid bedrock or till, thus introducing basal shear stress τ_b and basal velocity u_b and their interactions. In addition, basal sliding is correlated with cavity formation and size in the subglacial system, which can introduce feedbacks to the subglacial hydrology-sliding system. Understanding how ice slides across deformable and non-deformable beds has been a subject of much study in past decades, and there have been a number of mathematical analyses done to describe it.

The first sliding law was proposed by Weertman in 1957 [155]. Weertman considered a very idealized model of glacier beds in which bed roughness is represented by equal-size cubes that water flows around (Figure 4-4). His formulation considered two major concepts:

1. **Water film:** Assuming that for a temperate glacier, the temperature T is very close to the melting temperature of ice T_M, a film of water will form at the ice-bed interface due to frictional heating and pressure.

2. **Regelation:** Because the system is so close to T_M, water will melt on the upstream side of obstacles, where lots of frictional heating happens, and then re-freeze once it reaches the lower-pressure side of the obstacle.

However, there are many factors that contribute to the complexity of an accurate sliding law. The heterogeneity of glacial till means that a spectrum of obstacle sizes must be considered [52]. The non-Newtonian flow of ice introduces non-linearity, and if the glacier is temperate as most glaciers are, melting and refreezing of ice and water introduces a variety of mechanical and thermodynamic complexities. Since 1957, many studies have been done that have introduced additional levels of complexity to Weertman's simple model. In general, these laws follow the form $u = U(\tau)$ where u is the sliding velocity and U is some function of basal shear stress τ. Some parameters that can affect the sliding velocity $U(\tau)$ are subglacial water pressure, effective pres-

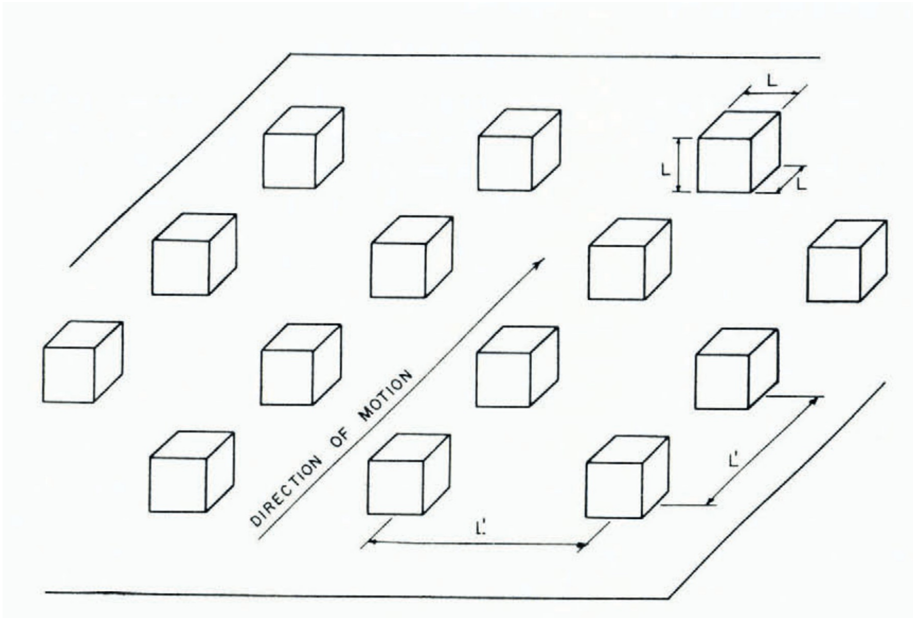

Figure 4-4: *An idealized illustration of bed roughness used in Weertman's hydrology based sliding model [155].*

sure, friction coefficient of the underlying bed, ice and water density, and thickness of the water film [52].

It is difficult to validate these models with real-world data because the basal conditions are rarely known [52]. However, there have been a few efforts to constrain these models further using experimental setups - for example, [161]. Understanding basal sliding is still a significant challenge in glaciology today.

Subglacial Hydrology and Sliding

Sliding of ice in temperate environments means that the bed temperature T is close to the melting temperature T_M, so we see a lot of melting and freezing. Subglacial hydrology models can provide good estimates of how much meltwater is being produced from dissipation of mechanical energy [146], and can account for other inputs of water such as rainwater and surface melt that makes its way to the bed. It has been observed that higher basal fluxes lead to higher velocities – for example, in seasonal and diurnal speedups observed in [61, 81, 141, 162]. It is also observed that there are higher sliding velocities near the termini of glaciers, where the bed elevation and hydraulic potential are lowest and subglacial water flux is high [81]. An intuitive representation of the relationship between water flux and the basal sliding coefficient was proposed by [59]:

$$A_b = A_0 \exp\left(\frac{\phi}{\phi_0}\right) \tag{4.5}$$

where A_b is the basal sliding coefficient, ϕ is the subglacial water flux, ϕ_0 is a limit factor on subglacial water flux, and A_0 is the initial value of A_b obtained from the nudging method [81]. This equation shows the basal sliding coefficient increases exponentially with subglacial water flux.

As mentioned earlier, the relationship between subglacial hydrology and sliding is also a complicated one. High flux regimes lead to a negative feedback loop whereby large fluxes lead to increased channelization and the formation of an efficient system that is able to shuttle water out of the system more quickly than it is supplied, thus ultimately decreasing the subglacial water pressure and slowing down ice velocity (discussed in 4.1.3). High subglacial water pressure has also been associated with glacial uplift and bed separation (for example, [72]), and phenomena such as till dilation bring in other feedbacks that have not been well understood thus far.

4.1.5 Differences between the Polar and Alpine Environments

While a great deal of in-situ observation and modeling work has been applied to the subglacial hydrology of the Greenland and Antarctic ice sheets, relatively less work has been done in alpine environments where human settlements lie close to glaciers - for example, the Andes and Himalayan mountain regions. These areas are usually very remote and are characterized by harsh and dangerous field conditions [150]. In addition, the Andes and Himalayan mountain regions are both contained within multiple sovereign nations, which makes it more politically difficult for international collaboration on in-situ data collection to occur.

A key physical difference between alpine glaciers and polar ice sheet glaciers is the slope of the bed on which the glacier rests. Although the land under polar ice sheets is quite heterogeneous, as a general rule most of them tend to be less steep than most alpine environments. This means that in mountain glaciers, hydraulic potential gradients are steeper. This has implications for how the subglacial hydrology develops. Röthlisberger channels typically dominate areas of high potential gradients - we therefore expect to see highly efficient, channelized drainage systems developing in alpine areas [50].

Because of this steep topology, alpine glaciers are more likely to be temperate and polythermal, meaning that some or all parts of the basal environment is raised to the pressure melting point for at least part of the year [67]. In addition, due to the steep topography, the top of the glacier (accumulation zone) may experience different temperatures than the bottom/terminus of the glacier near the ablation zone [67].

4.2 Observing Subglacial Hydrology

Subglacial hydrology is quite difficult to observe with the naked eye because the important systems are underneath glaciers and are physically difficult or impossible to get to. As a result, much of what we know about subglacial hydrology comes from remote sensing and indirect in-situ measurements.

4.2.1 Remote Sensing Methods

Remote sensing has been widely used by geoscientists since the 1970s. Remotely-sensed satellite data allows us to access high-resolution images with good temporal resolution even when physical access to a location is difficult or impossible.

Ice-Penetrating Radar

Ice-penetrating radar takes advantage of the differing dielectric properties of water and ice. Water shows a higher reflectivity of radar waves than both water and bedrock, and will clearly show up in radar cross-sectional images (Figure 4-5). For example, [34] used an ice-penetrating radar system to identify catchment-scale areas of englacial water storage (Figure 4-6). They found that the winter hydrological system in the

Greenland ice sheet was characterized by a linked-cavity system, whereby the cavities became isolated in the fall as smaller channels closed [34].

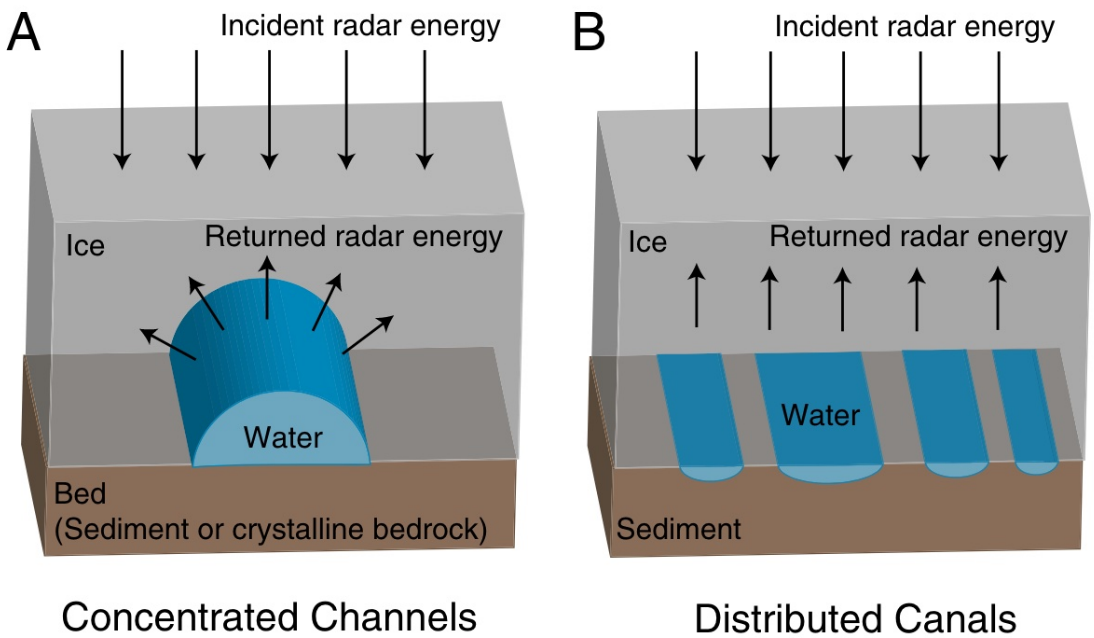

Figure 4-5: *How EM waves (radar) are scattered by channels vs. canals (i.e., efficient vs. distributed systems) - from [135]*

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ApRES (Autonomous Phase-Sensitive Radar) is a specific type of ice-penetrating radar that uses radar sounding echos to gather information about melt rates and water content in the glacial subsurface. ApRES systems are set up at the surface of the ice and measure the attenuation of transmitted radar waves [82]. Heat conduction, water content, and scattering all attenuate the ApRES signal and allow researchers to infer where englacial water is stored. For example, [82] found a 1-3m thick layer of surface meltwater stored in a macroporous layer in Store Glacier, Greenland based on scattering losses and radar attenuation from their ApRES deployment. They also found a periodic oscillation in bed echo power which allowed them to conclude that repeated drainage events were coming from the subglacial network [82]. ApRES is a very powerful method for understanding the subglacial hydrological environment; however, the machinery must be placed in the ice by hand during field campaigns,

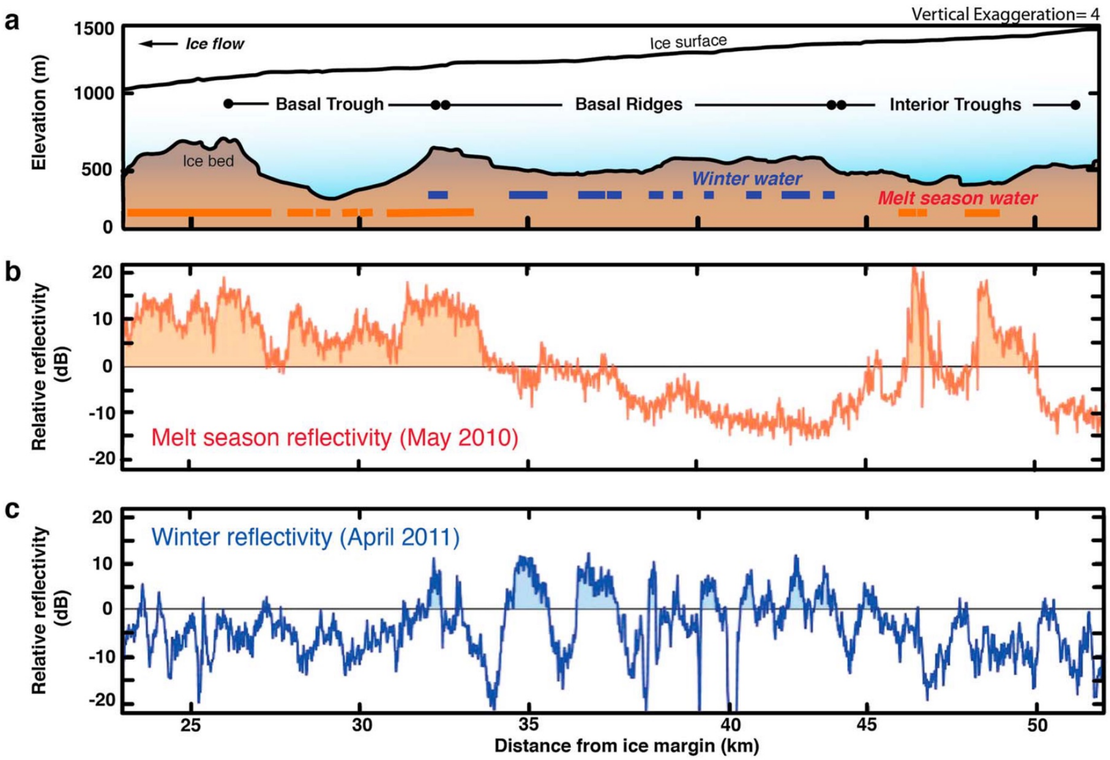

Figure 4-6: *Areas where englacial water is stored, such as in basal troughs and ridges, show higher radar reflectivity in ice-penetrating radar images [34].*

which can make it difficult to install them in remote areas.

4.2.2 In-Situ Methods

Boreholes and Dye-Tracing

Drilling boreholes is one of the oldest methods for studying glaciers and subglacial hydrology. Boreholes are generally used to assess water pressure at the bed in various locations, and are the only direct in-situ observations we have available [50]. The disadvantage of boreholes is that they only offer point-scale observations and can be quite expensive to drill, meaning that they do not provide the scale required to develop a complete understanding of subglacial systems [50]. Rather, borehole measurements are used to calibrate models and other catchment-scale observations of subglacial hydrology.

Dye-tracing provides another way to analyze how water travels through the subglacial network. Typically tracers are made of fluorescent dyes, which are injected into a glacier on the surface via a moulin (e.g., [138]). Water samples are continuously taken at the drainage area at the glacier terminus and are tested using fluorometry to determine the residence time of the tracer inside the glacier [33]. For glaciers where the residence time is expected to be high, different tracers can be used, such as sulfur hexafluorine gas (SF_6) which was used in [33] in their study of the Greenland ice sheet. Dye-tracing studies have been carried out in a variety of alpine glacier environments, including in the Nepal Himalaya [98]. By analyzing the residence time and dispersion of the dye (measured at peaks of dye concentration) the characteristics of the drainage system can be inferred [98].

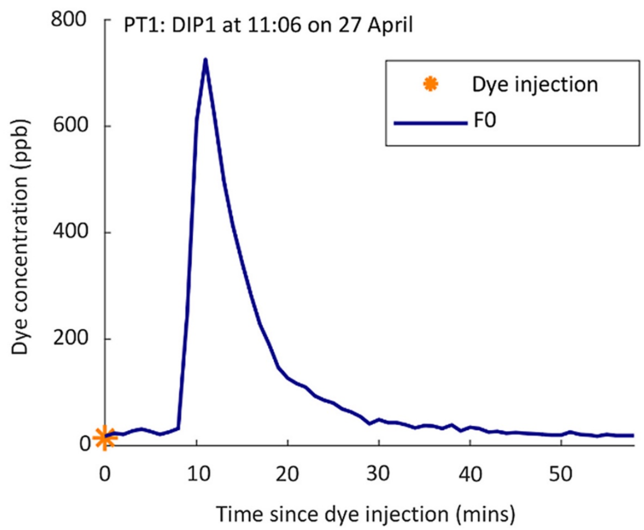

Figure 4-7: *Concentration of fluorescent dye at the terminus of Khumbu Glacier, Nepal as measured by [98].*

Discharge Analysis

The bulk discharge from a glacier snout can be analyzed to shed light on englacial water storage and residence time. According to [67] water that is stored within a glacier for longer periods of time is ionically enriched due to weathering from the drainage system - for example, Ca^{2+}, Mg^{2+}, SO_4^{2-} and Si can commonly be found in water that has passed through a subglacial water network. Discharge measurements have also been combined with seismic measurements as part of mathematical inversions to determine the characteristics of hydraulic pressure gradients in the subglacial system [108].

Geophysical Methods

In-situ global positioning system (GPS) measurements are used by many studies to track the velocity of ice through time [21, 141, 35]. For example, [21] set up four GPS trackers on an ablation zone on the Greenland ice sheet to track diurnal and seasonal changes in velocity that shed light on the state of the subglacial hydrology system. GPS systems can also provide high-precision elevation measurements; [90] used GPS to track the elevation of ice blisters over time to infer the transmissivity of the subglacial drainage network from the relaxation time of the blisters following supraglacial lake drainage. The Global Navigation Satellite System (GNSS) similarly uses satellites to track surface velocities.

Seismic monitoring can also provide insight into the hydrology under glaciers. Abrupt movements and drainages can cause increases in seismic activity and are detected by seismometers. For example, [86] used passive seismic monitoring to detect rapid lake drainage and subsequent uplift in Kangerlussuaq, Greenland. They were also able to fit curves to the data to determine the material underneath the glacier based on its seismic response to the lake drainages. Seismometers must be placed manually in the field, and thus can be difficult to use in remote or inaccessible areas.

Geomorphological Observations

A final tool that is often used to understand subglacial hydrology is observation of how previously glaciated landforms have been shaped by the glaciers that used to exist there. In the lowlands of Northern Germany, tunnel valleys, which are hundreds of meters deep, indicate evidence of subglacial, sediment-controlled canals which exhibit an anastomosing pattern [47]. Although the glaciers that once covered these landforms are gone, the canals can offer insight into how those glaciers might have looked like.

4.3 Modeling Subglacial Hydrology

Efforts to model subglacial hydrology began in the 1970s. The first subglacial hydrology models was conceived by Weertman in 1972, who proposed a thin, sheetlike film of water flowing at the base of an ice sheet or glacier [156]. Röthlisberger's 1973 work talked about channels cut into the ice at the base of the glacier, imagining a subglacial drainage system that is in equilibrium between melt opening and creep closure (plastic deformation) [130]. More advanced models have added on complexity, describing the subglacial water system as a network of connected channels and linked cavities [80]. Today, subglacial hydrology models generally establish that the system can be classified either as efficient or inefficient [146]. The difference between these two classifications is expanded upon below and in Figure 4-1.

Efficient Systems An efficient drainage system is characterized by the presence of channels and canals which are able to quickly and "efficiently" transport water from place to place. This idealized system is also referred to in the literature as "fast" or "channelized" [50].

Inefficient Systems The inefficient drainage system is characterized by the flow of water through flat, distributed geometries such as sheets, cavities, and pore spaces [50]. This type of network system is also referred to as "slow" or "distributed".

Subglacial hydrology models require a few essential elements [50]:

1. **Conservation of Mass:** Under the assumption that water is incompressible, the amount of water in the system must remain constant. This is expressed in the general form shown below, where h is water volume or depth, q is water flux, and b is a source/sink term.

$$\frac{\partial h}{\partial t} + \nabla \cdot q = b \tag{4.6}$$

2. **Water Sources and Sinks:** The total amount of water in the subglacial system is affected by a variety of processes. These can be linearly combined using simple addition of the form:

$$b = b_s + b_e + b_a + b_b \tag{4.7}$$

where b_s is surface runoff from melt and rain, b_e represents englacial processes (e.g., strain heating), b_a are from groundwater discharge/recharge and aquifers, and basal production and consumption is b_b [50]. Discharge to rivers and glacial lakes can also be a factor here.

3. **Conservation of Linear Momentum:** Conservation of momentum gives us a way to calculate flux and velocity of the water. The general form of momentum conservation is expressed as

$$q = -Kh^{\alpha}(\nabla\phi)^{\beta} \tag{4.8}$$

where K is a rate factor, $\nabla\phi$ is the fluid potential gradient, and α and β are parameters describing the drainage system including its efficiency and Reynolds number of the flow [50].

4. **Driving Force for Water Movement:** Differences in potential energy drive the movement of water through the subglacial hydrological network. Water travels "down" gradients, from areas of high potential energy to low potential energy. Potential energy gradients are expressed by differences in hydraulic head, which combines water pressure and gravitational potential energy.

4.3.1 SHAKTI: Subglacial Hydrology and Kinetic, Transient Interactions

In Chapters 5 and 6, we will employ the SHAKTI model to simulate the subglacial hydrology of an alpine glacier. SHAKTI is a numerical model first developed in 2018 by Sommers et al. [146]. Unlike its predecessors in subglacial hydrology modeling, SHAKTI is able to model a variety of network systems between the end-member cases of efficient and inefficient drainage systems [146]. It does this by allowing the hydraulic transmissivity to vary spatially and temporally [146]. In addition, the model is able to account for varying laminar, turbulent, and intermediate regimes [146]. An additional benefit to this model is that it can be coupled with ISSM (Ice-Sheet System Model) which allows us to model ice sheet dynamics and glacier flow along with the subglacial hydrology - something that had not been done extensively in the past.

Model Formulation

The model is formulated with several equations and conditions that are applied to the entire domain. The subglacial environment is modeled as a continuous sheet of differing gap height, and the gap height is altered through the processes of melt opening (positive) and creep closure (negative). Channels are represented by chains of large gap height, and cavities are opened by bumps with user-prescribed spacing and size.

The following equations are used in the SHAKTI model:

1. **Water mass balance**, analogous to Equation 4.6, is modeled as the following:

$$\frac{\partial b}{\partial t} + \frac{\partial b_e}{\partial t} + \nabla \cdot q = \frac{\dot{m}}{\rho_w} + i_{e \to b} \tag{4.9}$$

   where b is the subglacial gap height, b_e is the volume of water stored englacially per unit area in the bed, q is basal water flux, $\dot{m}$ is basal melt rate, and $i_{e \to b}$ is the rate of flow from the englacial to subglacial system. Similar to Equation 4.6, water is assumed to be incompressible. The model also assumes here that the subglacial gap space is always filled with water.

2. **Evolution of gap height**, which represents the subglacial geometry, is expressed as

$$\frac{\partial b}{\partial t} = \frac{\dot{m}}{\rho_i} + \beta u_b - A|p_i - p_w|^{n-1}(p_i - p_w)b \tag{4.10}$$

where β is a dimensionless parameter controlling opening by sliding, A is Glen's ice-flow law parameter, p_i is the ice overburden pressure, and p_w is water pressure. The first term in the equation above represents melt (turbulent) opening, the second term represents opening by sliding, and the third term is creep closing.

3. **Horizontal basal water flux** q can be represented by

$$q = \frac{-b^3 g}{12\nu(1 + \omega Re)} \nabla h \tag{4.11}$$

where g is gravitational acceleration, ν is kinematic viscosity of water, ω a dimensionless parameter controlling the transition from laminar to turbulent flow, h is hydraulic head, and Re is the Reynolds number $Re = \frac{|q|}{\nu}$.

4. **Internal melt generation** occurs as a result of geothermal heat flux, frictional heating from sliding, and internal dissipation. It is expressed by

$$\dot{m} = \frac{1}{L}(G + |u_b \cdot \tau_b| - \rho_w g q \cdot \nabla h - c_t c_w \rho_w q \cdot \nabla p_w) \tag{4.12}$$

where L is is the latent heat of fusion of water, G is geothermal flux, u_b is the ice basal velocity vector, τ_b is the stress exerted by the bed onto the ice, c_t is the change in the melting point caused by pressure, and c_w is the heat capacity of water. The last term provides an adjustment for the energy consumed or released to maintain the temperature of water at its pressure melting temperature. The model neglects heat advection and assumes that the water is always at the pressure melting temperature.

The above equations are combined to obtain a differential equation that can be uniformly applied across the model domain:

$$\nabla \cdot (-K \cdot \nabla h) + \frac{\partial e_v(h - z_b)}{\partial t} = \dot{m}\left(\frac{1}{\rho_w} - \frac{1}{\rho_i}\right) + A|p_i - p_w|^{n-1}(p_i - p_w)b - \beta u_b + i_{e \to b} \tag{4.13}$$

where K is the hydraulic transmissivity tensor $K = \frac{b^3 g}{12\nu(1+\omega Re)} I$. This is a non-linear, elliptic partial differential equation. It can also be modified to include an englacial storage parameter, e_v. By default, the englacial storage parameter is set to 0 such that surface water reaches the ice-bed interface instantaneously. However, a spatio-temporally varying englacial residence time can be assigned across the mesh.

Major Assumptions

There are a few major assumptions that SHAKTI uses that must be considered when analyzing model results:

The underlying bed is non-deforming and made of solid bedrock. We have discussed the difference between hard and soft beds in 4.1.2. Using this assumption means that the model does not account for sediment transport/erosion, till dilation, water storage in sediments, and channelization via sediment creep.

Any surface meltwater reaches the bed instantaneously. We have seen in 4.1.1 that although there is a strong coupling between surface melting and delivery to the bed, storage in englacial aquifers can complicate this relationship. We can get around this by manually adding residence times in englacial storage. However, this flux is not directly modeled by SHAKTI.

The model disallows water pressure to be higher than the ice overburden pressure (i.e., negative effective pressure), which prevents uplift of ice or bed hydrofracturing. Ice uplift due to negative effective pressure has been observed in areas and times of very high water pressure, but are complex phenomena that are not yet considered in SHAKTI which is a 2D model.

Even with the limitations posed by these assumptions, SHAKTI is at the cutting edge of subglacial hydrology models. Due to its ability to model a range of viscous regimes (laminar to turbulent) and types of drainage systems (channelized to distributed), it allows us to understand the seasonal evolution of the subglacial drainage system.

Coupling with ISSM

SHAKTI is implemented as part of the open-source Ice Sheet System Model (ISSM), one of the largest, most comprehensive, and commonly-used ice sheet models today. It was developed by NASA's Jet Propulsion Laboratory and the University of California at Irvine [91]. ISSM includes a variety of packages that allow for the modeling of ice flow, calving, thermal modeling, grounding line movement, damage mechanics, glacial isostatic adjustment (GIA), stress balance, and more [91]. The coupling of ISSM and SHAKTI gives the potential to model a variety of very complex glacial phenomena.

4.4 Chapter Conclusion

In this chapter, we discussed a variety of topics in subglacial hydrology that will prove useful for the work that will be undertaken in Chapters 5 and 6 on surging and GLOFs. First, we discussed how to characterize the subglacial drainage network. We saw the journey surface meltwater takes to get to the ice-bed interface. We then looked at the channelized vs. linked-cavity systems that could form in response, and the relationships between effective pressure and flux that then emerge in each of these types of systems. We also considered how the case of a deformable bed may complicate these feedbacks. We then built on these ideas of meltwater delivery and channelization to understand how the drainage system changes across seasons. We learned that spring and summer fluxes cause the system to become more efficient and channelized, whereas the decrease in flux in the fall and winter allow the system to return to an inefficient, distributed system. We then investigated the relationship between system efficiency and basal water pressure, and did a brief exploration of the sliding laws that then determine the basal velocity of the glacier.

In the second section, we briefly discussed observational methods for studying subglacial hydrology. This included remote sensing methods such as ice-penetrating radar, which have revolutionized the field of subglacial hydrology. We also looked at in-situ methods such as borehole drilling, dye-tracing, discharge analysis, GPS, GNSS, seismic monitoring, and geomorphological observations. For each of these methodologies, we took a look at a case study that used those techniques to learn something new about the subglacial system.

In the final section of this chapter, we discussed a major frontier of subglacial hydrology, and one that we will focus on the most in coming chapters – modeling. We discussed how different subglacial drainage regimes are modeled as "efficient" or "inefficient." We took a look at the essential pieces of a subglacial hydrology model. Lastly, we introduced the SHAKTI model, which we will employ in the research section of this thesis. We talked about SHAKTI's strengths relative to previous subglacial hydrology models and showed the most pertinent elements of the model formulation. We also went over some of SHAKTI's major assumptions and how they could affect the model results.

We have now laid out the conceptual foundation for much of the work in the following chapters. Despite the large amount of knowledge we do have about subglacial hydrology, it is a field that is rife with uncertainty, data scarcity, and unknowns. In the next chapters, we will use the concepts we learned in this chapter to glean insights into the behavior of a glacier in northern Pakistan, Shishper Glacier, which has exhibited both surge behavior and contributes to glacial lake outburst flooding.

Chapter 5

Case Study: Modeling of Subglacial Hydrology in Gilgit-Baltistan, Pakistan

5.1 Introduction

The Himalaya-Karakoram mountain region (High Mountain Asia) is known as the "Third Pole" due to its immense amount of glacial ice mass. The glaciers of High Mountain Asia feed major water systems which provide water and sanitation for over 1 billion people [9]. In particular, the Karakoram is the most heavily glaciated range in Asia [39] and is a critical water source for the country of Pakistan. However, with climate change, mass balance has been decreasing, putting the future of the area at risk. Glacial hazards like glacial lake outburst floods (GLOFs) are also expected to increase in severity and frequency, causing widespread damage. Understanding the role subglacial hydrology plays in these hazards can help us develop early-warning models for water availability and outburst flooding. Given the relative lack of observational data in High Mountain Asia, modeling provides a way to bridge the gap of understanding. In this work, we use a new, unified subglacial hydrology model SHAKTI to model the seasonal evolution of subglacial drainage in Gilgit-Baltistan, Pakistan.

Shishper and Muchuwar

The study site for this research is Shishper Glacier (36.40°N 74.61°E) in the eastern Karakoram range in Pakistan (Figure 5-1). It is located in Hunza Valley in Gilgit-Baltistan, Pakistan. Shishper is part of a surge and drainage system with another glacier to its west, called Muchuwar. The two glaciers used to be connected prior to 1950, when the two separated [103]. Shishper's main trunk is approximately 7 km long and is fed by several tributary glaciers at the northeast (upper-elevation) side. In total, the glacier is about 15 km at its longest [23]. The site was chosen due to its surge-type behavior and history of lake drainages (GLOFs). Additionally, there have been several previous studies on the area which can provide data to support and compare models results with [26, 103, 122, 23].

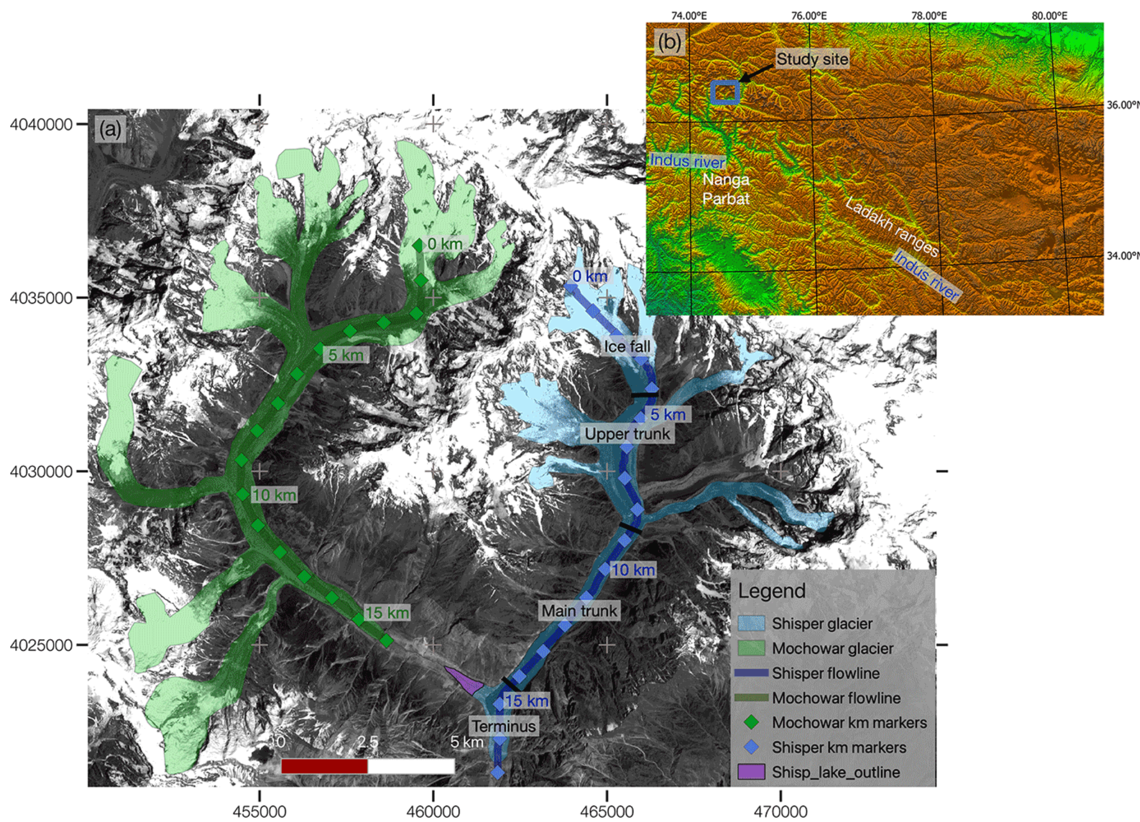

Figure 5-1: *Overview of Shishper (blue) and Muchuwar (green) Glaciers in Gilgit-Baltistan, Pakistan - from [23]*

Both Shishper and Muchuwar have been subjected to cyclical surges for as long as observations have been recorded, since the early 1900s [23]. Shishper underwent a major surge between 2017-2019 and also in 2000-2001 and 1973 [26]. During this time, Shishper advanced approximately 1.5 km [26]. The surge resulted in the closing of two power plants and the evacuation of nearby villages. The movement of Shishper blocked meltwater flow from Muchuwar Glacier, which created an ice-dammed lake which we will hereafter refer to as Hassanabad Lake (outlined in purple in Figure 5-1). The lake tends to fill up in November-December and in May to a depth of 30-80m, with an estimated volume of 30 million m^3 [149]. When the lake drains, the outburst flood drains through the terminus of Shishper and down into the valley below. The maximum river flow observed at the downstream village of Hassanabad is 150-200 m^3 s^{-1}, compared to a base flow of about 20 m^3 s^{-1} [149]. After the lake is filled in the winter, drainage occurs more gradually, as opposed to the spring filling which results in a more dramatic drainage of the lake.

A devastating GLOF was most recently recorded in May 2022, resulting in the destruction of part of the Karakoram Highway, and the displacement of about 20 residents of Hassanabad [20]. The Karakoram Highway is the only road connecting Pakistan and China. [144] also suggest that due to warming summer temperatures in the area, another glacial lake may form at the terminus of Muchuwar Glacier, which has the potential to create cascading effects with Shishper Lake in the future. The availability of previous studies on this lake and its behavior of both surging and GLOFs, it provides a data-rich site for modeling.

5.2 Model Setup

As mentioned earlier, modeling of subglacial hydrology is an important method for improving our understanding of water flow underneath Himalayan glaciers, especially given the lack of in-situ measurements in the H-K region. In this work, we use the model SHAKTI: Subglacial Hydrology and Kinetic, Transient Interactions [146], which is implemented as a part of the Ice-Sheet System Model (ISSM). Due to its ability to simulate a variety of flow and drainage regimes, as well as the potential to couple the subglacial hydrology with ice flow dynamics, SHAKTI was the clear model choice for this work. The advantages and disadvantages of the SHAKTI model, as well as the assumptions it is based on, were discussed in Section 4.3.1.

The model domain was set up using the following remotely-sensed inputs:

Glacier Outline was taken from the Randolph Glacier Inventory (RGI) Version 6.0 [39]. The glacier outlines contained in this dataset are from remotely sensed images recorded in 2016. The GLIMSID for Shishper Glacier is G074619E36437N.

Surface Elevation was from 90-m TanDEM-X global Digital Elevation Model (DEM). (Figure 5-2a).

Bed topography Bed topography was inferred from surface elevation and bed thickness, which was obtained from [49]'s global glacier thickness dataset, which is publicly available online. Glacier thickness here is calculated using an inversion from ice flow dynamics.

Surface Velocity Surface velocity is obtained from ITS_LIVE [2], a NASA product which provides velocity measurements at 120m spatial resolution for glaciers across the globe.

Since the tributary glaciers are at much higher elevation than the main trunk, they likely do not experience as much liquid precipitation. There is also likely not as much melt generation at these elevations because the bed temperature is much colder there. Although the tributary glaciers may contribute to the mass accumulation of the glacier, their contributions to the subglacial hydrological system are probably less important. Therefore we removed the higher-elevation tributary glaciers from the model domain (Figure 5-3). Modeling the contributions of the tributary glaciers on the subglacial system may be part of future studies as we add complexity to the model.

The outline of the domain is drawn manually to the RGI outline using functionalities of ISSM in MATLAB. Remote sensing imagery was processed and projected to

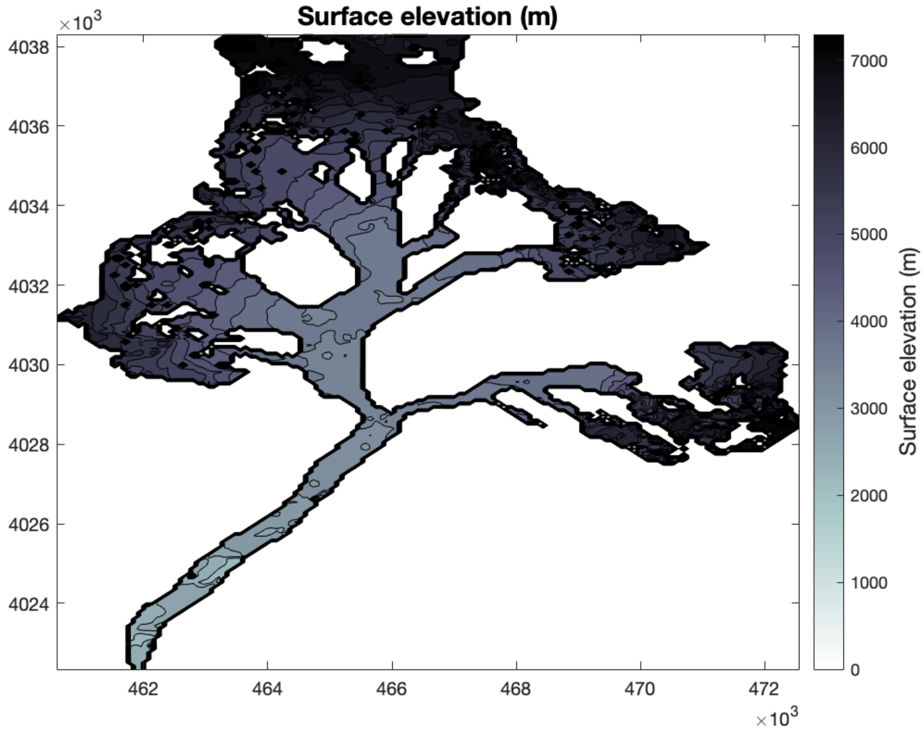

Figure 5-2: *A contour plot of the surface elevation of Shishper Glacier, from the TanDEM-X global DEM.*

the 43N UTM (Universal Transverse Mercator) coordinate system using QGIS. These data are interpolated onto an unstructured triangular mesh with 50m resolution. This mesh provides the basis for the P1 triangular Lagrange finite element solver used by SHAKTI. All simulations in this work are carried out with ISSM Version 4.23 using the MATLAB interface on MacOS.

5.3 Preliminary Simulations

The eventual goal of this project is to run transient simulations with continuous changes in meltwater input across seasons. However, transient simulations are computationally expensive and require a good intuition for an appropriate timestep, as SHAKTI does not yet have adaptive timestepping built in. We also need to calibrate the model's initial parameters. To do this, we run a few preliminary simulations for the primary purpose of informing the transient simulations, which will be discussed

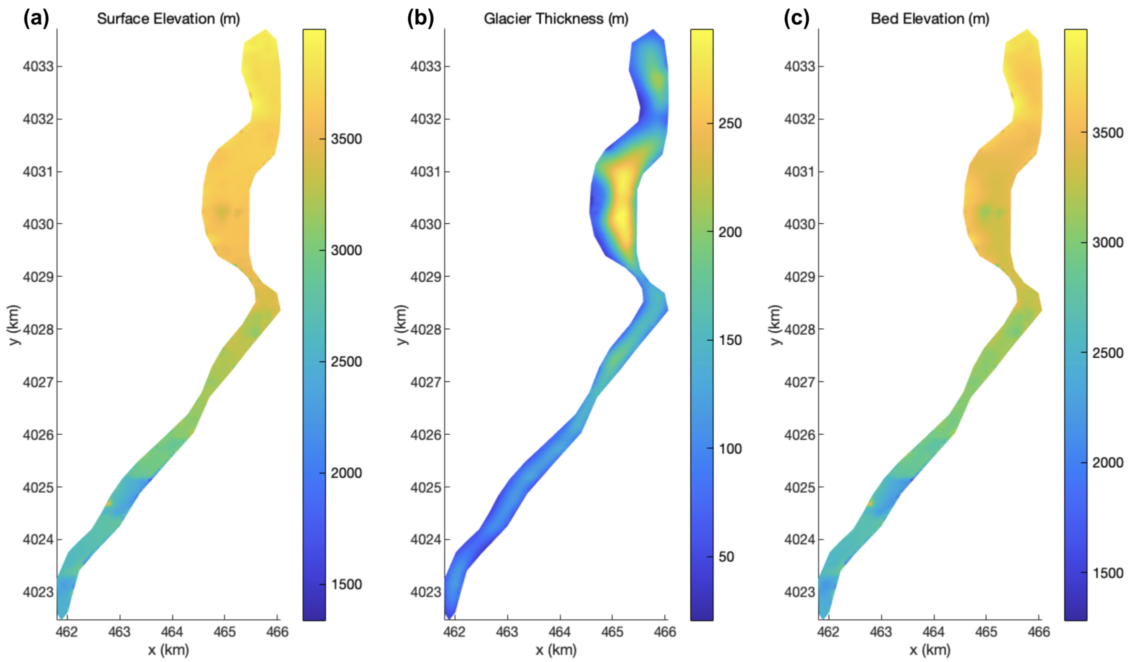

Figure 5-3: *(a) Surface elevation,(b) glacier thickness, and (c) bed elevation over the modeled domain of Shishper Glacier.*

in Section 5.4. These preliminary simulations can also help us better understand the behavior of the model under a few simplified, steady-state conditions. These model runs can also serve as a "sanity check" to make sure the model is behaving as we expect it to.

It is important to note that the models shown here are uncoupled with ice dynamics; that is, the surface and basal velocity were prescribed as 0 across the mesh and across time. While velocities of 0 are unphysical in the real world, especially during melt season, the goal of our first forays into modeling this glacier with SHAKTI is to understand how the subglacial drainage system itself will change seasonally. In previous chapters, we have discussed the feedback systems between the subglacial drainage network and ice flow; however, these feedbacks are not included in these models. This provides us a unique opportunity to gain a true understanding of the hydrological system without the interference of ice flow dynamics.

5.3.1 Winter Spin-Up

Before seasonal simulations can be run, the base winter state of the hydrological system must be established. To do this, we prescribe arbitrary initial input parameters and allow the system to equilibrate to its base state. During the winter, we assume that there is no surface or englacial melt; the only hydrological input to the bed is through geothermal flux, and melt opening/turbulent dissipation (note that we have prescribed sliding velocity to be 0, so there is no frictional heating). The initial parameters for the 200-day spinup are shown in 5.1. Because we are excluding all contributions from tributary glaciers, a Neumann boundary condition of zero flux is applied to all edges of the domain. A timestep of 1 day is used for the spinup; although relatively coarse, the winter spin-up does not require finer temporal resolution because we are only trying to obtain the final, equilibrated state.

Parameter	Prescribed Initial Value
Hydraulic head	50% of ice overburden
Gap height	0.001 m
Bump spacing	2 m
Bump height	0 m
Meltwater input to bed	0 m/yr
Reynolds number	1000
Flow law exponent	n=3
Velocity	0

Table 5.1: *Initial values prescribed to model parameters for the winter spin-up run.*

From trial and error, we found that the system took approximately 200 days to fully reach steady state. Figure 5-4 shows that all parameters except for gap height reach equilibrium between 150 and 200 days of simulation. In future work, we will run the winter spin-up longer so that gap height reaches its equilibrium state; for now, this is acceptable to us because the gap height is not growing exponentially.

Once the model reaches equilibrium, there is clear formation of a channel down the main trunk of the glacier (Figure 5-5). In SHAKTI, channels are modeled as chains of higher gap height; the channel in Figure 5-5a, shown in yellow, is clearly corroborated by an opening of gap height (Figure 5-5b) which accommodates this

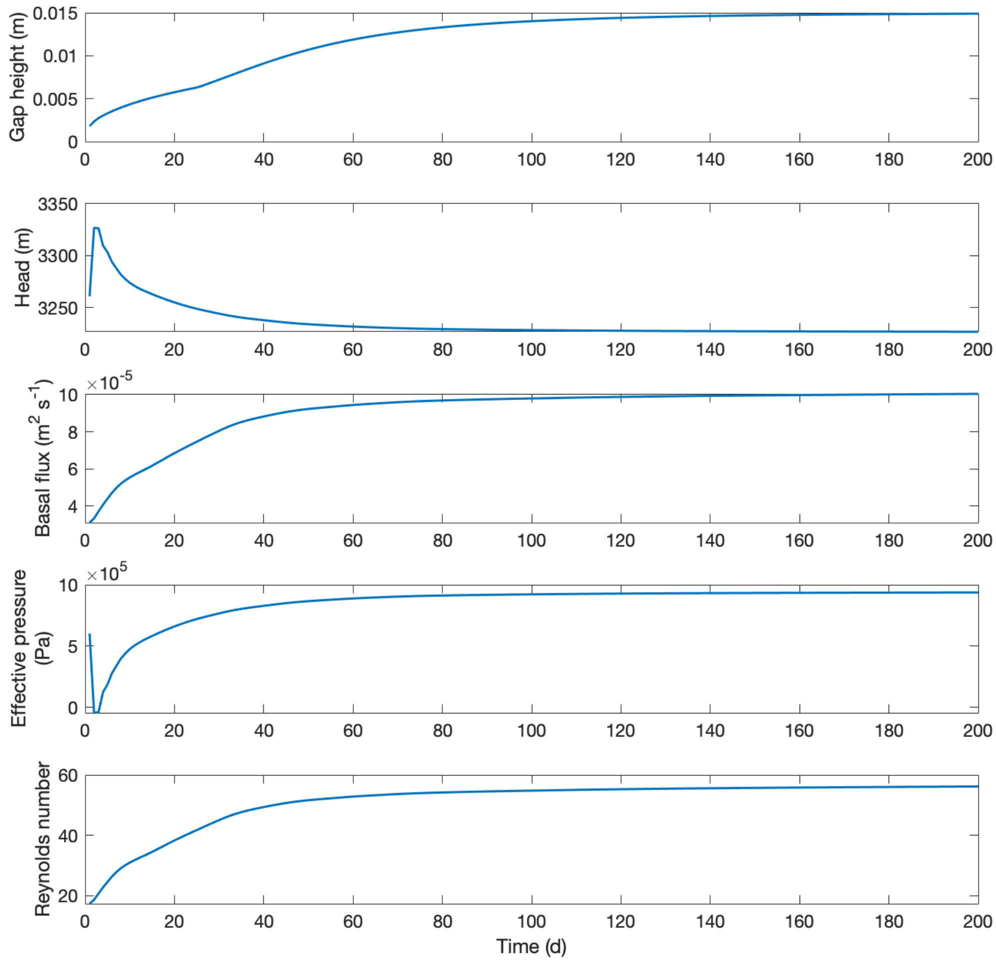

Figure 5-4: *Mean values of gap height, hydraulic head, basal flux, effective pressure, and Reynolds number across the mesh for the 200-day winter spinup.*

increased flux. It is important to note that the channel has formed in the absence of any surface water melt, indicating that it likely persists through the winter months and forms the main conduit of the subglacial system during the melt season.

In addition, according to [20], surging glaciers like Shishper often have subglacial (beneath the ice) lakes in their ablation zone, from which drainages can also occur. We observe two such lakes near the terminus in our model of Shishper. This gives us additional confidence in the model's fidelity to DEM and ice thickness data.

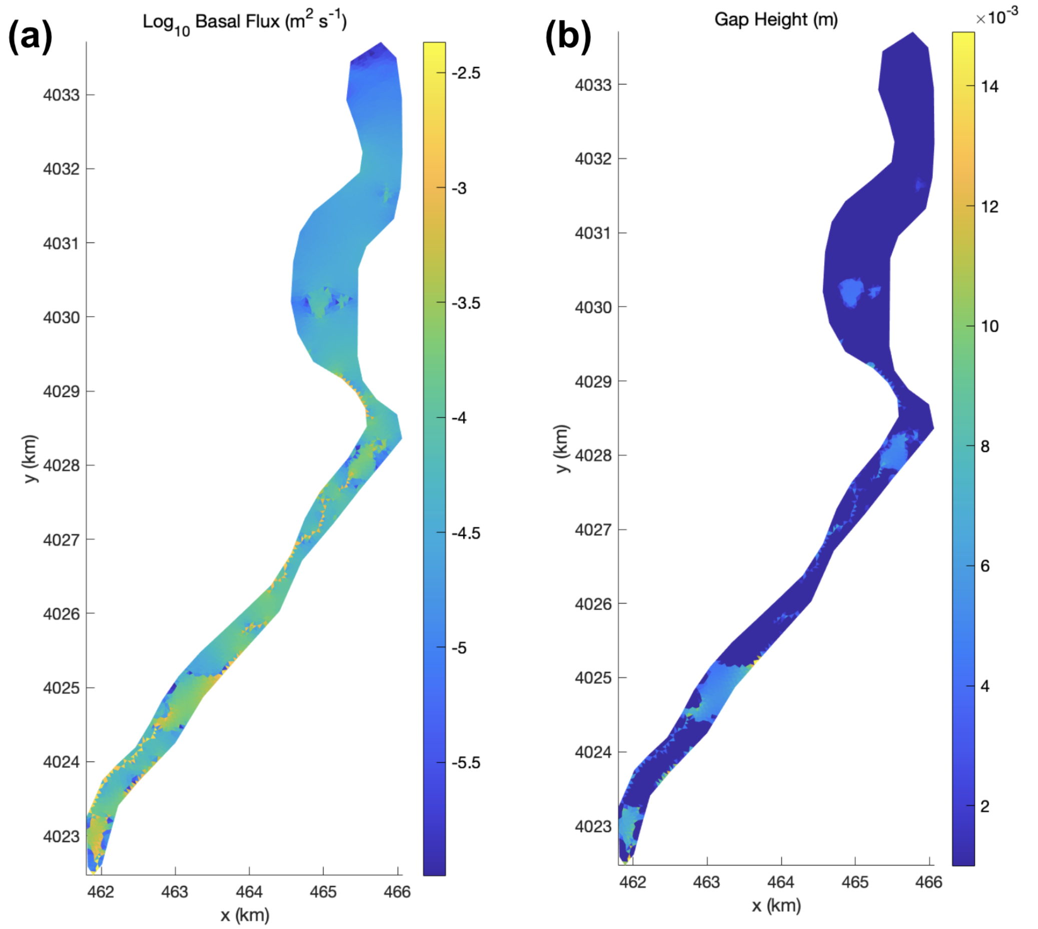

Figure 5-5: *(a) Basal flux and (b) gap height over the domain following the 200 day winter spin-up.*

5.3.2 Discrete Meltwater Inputs

To explore the effects of rising water quantities in the system, we performed several simulations at discrete hydrological inputs to the subglacial system (hereafter referred to as just "englacial input"). Rather than a continuous rise in englacial input as is seen in physical systems, we modeled a series of discrete steps with increasing melt. Starting with the basal conditions reached at the end of the winter spin-up, we added uniform, distributed melt inputs across the mesh. The englacial input is instantaneously delivered to the bed, and the simulations were allowed to run until the system reached equilibrium as visually determined by time-series plots like the ones in 5-4. The parameters for the seven model runs are shown in Table 5.2.

Englacial Input (m/yr)	Timestep (h)	Time until equilibrium (d)
1	3	20
2	3	20
3	3	20
4	3	30
5	2	50
6	2	50
7	0.5	45

Table 5.2: *Discrete englacial inputs to the winter system, the minimum timestep required for numerical stability, and the number of days for the system to reach steady-state.*

For each discrete input, we attempted the largest possible timestep, since runtime is directly proportional to the number of timesteps required for the model run. Generally, we saw that increasing englacial input to the subglacial system led to decreased numerical stability, which we accommodated by decreasing the timestep. This makes sense because we would expect larger water inputs to have a faster and more dramatic effect on the subglacial system; therefore, in order to maintain the continuousness of the model, smaller timesteps must be allowed. We also noted that the system took longer to reach steady-state in response to higher meltwater inputs. This makes intuitive sense as well, because we would expect the system to require more reorganization time given a more extreme change in input conditions.

The system undergoes various changes when we increase the distributed englacial input across the mesh. Figure 5-6 shows the differences in horizontal basal flux and hydraulic head from simulations with englacial input 1 m/yr to 7 m/yr. In Figure 5-6a (basal flux), most of the domain shows a difference of 0 m/yr between the two plots. However, in the central channel, the flux increases slightly. The growth of the channel's size (Figure 5-6c) accommodates an increase in flux. We also observe lots of changes in hydraulic head (Figure 5-6b); when more water is added to the system, we see a large decrease in head at the upper part of the glacier (dark blue) with smaller decreases at the middle of the glacier and a small increase in head (yellow-green) at the outlet. This shows us that the channel is efficiently transporting high-pressure

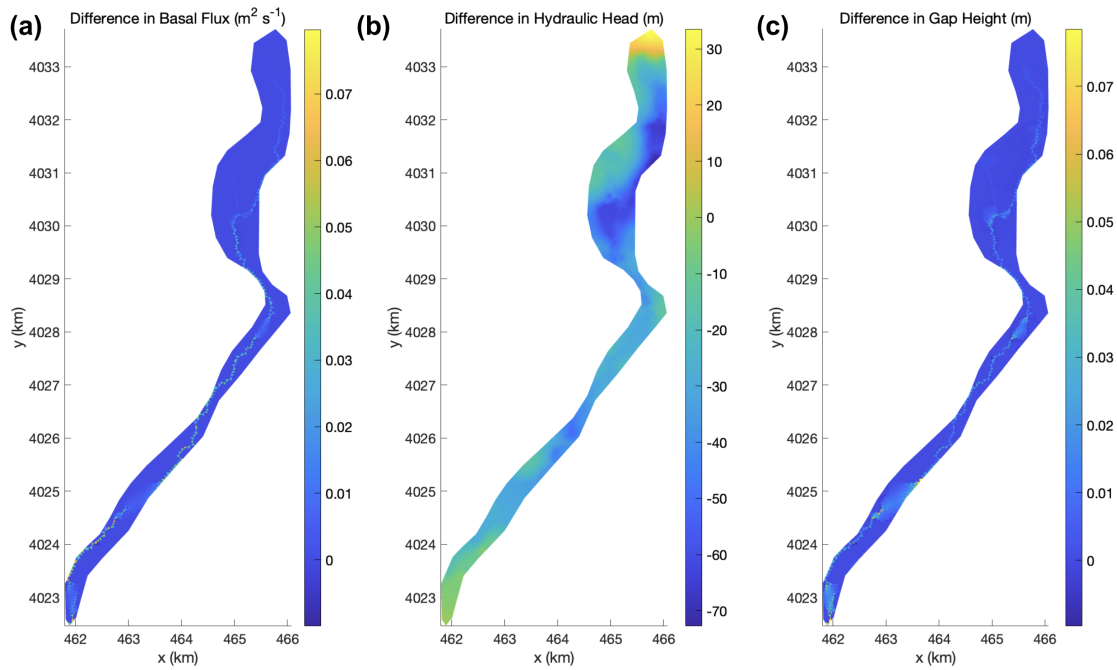

Figure 5-6: *Change in (a) basal flux, (b) hydraulic head, and (c) gap height between models equilibrated to 1 m/yr and 7 m/yr englacial inputs.*

flow from high to low gravitational potential. There is also a very large increase in hydraulic head at the very top of the glacier (bright yellow). Since the drainage channel does not yet reach this area, it is not able to evacuate the high levels of melt input that are entering the bed there, and water pressure/hydraulic head builds up.

A key takeaway from these preliminary simulations is that the subglacial drainage system *reorganizes in response to the amount of basal flux.* In Figure 5-7a, we see that as englacial input increases, basal flux increases at the outlet and middle of the glacier. However, the basal flux at the upper end of the glacier decreases slightly. This likely happens because greater englacial inputs cause more reorganization of the drainage system, as noted above. When we prescribe a higher englacial input, the system is more strongly forced to find ways to transport flow from high to low gravitational potential energy. By the time the system equilibrates to the prescribed input, there is more flux at the bottom of the glacier than the top.

The transmissivity of the drainage system, seen in 5-7b, supports this conclusion. Transmissivity, K is a material property of the system that tells us how easy it is

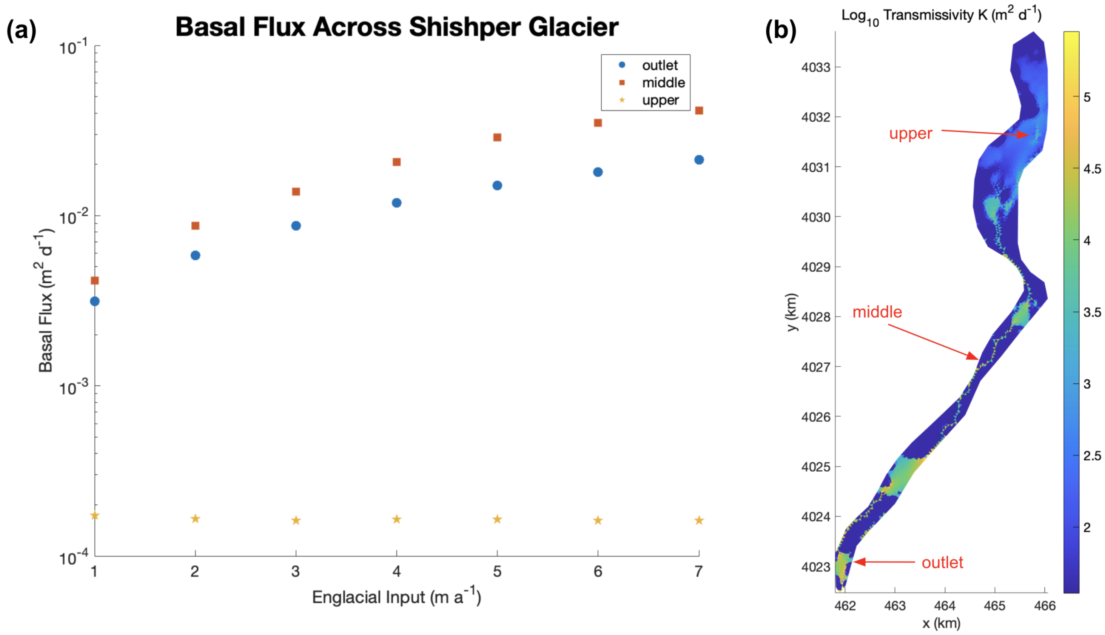

Figure 5-7: *(a) Basal flux (m^2 d^{-1}) for a single mesh triangle in the outlet, middle, and upper parts of Shishper's trunk vs. englacial input. (b) log_{10} of transmissivity K (m^2 d^{-1}) after equilibrating to a distributed englacial input of 3 m/yr. Transmissivity is shown in log scale to better illustrate differences across the domain.*

for water to flow through the system at that place (K is defined in 4.3.1). Since the transmissivity is low in the upper section of the glacier, we can infer that less reorganization of the drainage system has occurred there, leading to locally inefficient, distributed system. This lower-transmissivity area can be seen at the arrow labeled "upper" in Figure 5-7b; there is no distinct channel to be seen, and there seems to be a "blob" of low-transmissivity. At the middle and lower parts of the trunk, a single, narrow channel with high transmissivity and high horizontal basal flux forms, indicating that the system there has equilibrated to a channelized, efficient system. From our discussion in 4.1.3, this pattern agrees with our expectation that when basal flux is higher, more efficient, channelized systems form due to the dominance of the melt opening term. This result is encouraging because it shows agreement with past observations and models of the evolution of subglacial networks.

5.4 Transient Simulations

To better understand how Shishper's subglacial drainage network responds to seasonal changes in meltwater flux, we employ transient simulations across a period of five years. We chose the calendar years 2017-2021 to model the system conditions before, during, and after the surge event that occurred between late 2017 and mid-2019. The transient input for these simulations is *hydrological input to the bed*, which we prescribe to be spatially and temporally varying surface ice melt.

We do not include liquid precipitation in these experiments. The role of external events (i.e., sudden influx of water from storms) in triggering surge events is particularly uncertain [27]; therefore, simulating the ice melt input with and without liquid precipitation data may provide a unique insight into the role of storms in surge behavior. Future work (6.2) will elaborate on this further.

5.4.1 Melt and Precipitation Data Acquisition

Data from the European Centre for Medium-Range Weather Forecasts (ECMWF)'s Reanalysis v5 (ERA5) were run through a temperature-indexed ice melt model [93] to obtain spatio-temporally varying estimates for surface melt across the domain. These ERA5 weather data are based on an array of field stations and weather models [139], and directly provide estimates for snow melt, ice melt, and liquid precipitation across the five years. Ice melt across the mesh is calculated using the temperature-indexed (TI) melt parametrization developed in [93], which uses ERA5's air temperature, cloud cover, and calculated net shortwave radiation to estimate ice ablation. The TI model is shown to be more accurate for glaciers below 3500 m above sea level (a.s.l) – most of Shishper's main trunk lies below 3500 m a.s.l. (Figure 5-2), so this data provides suitable estimates for our use [93]. Finally, the ERA5 data is validated by data from the Pakistan Water and Power Development Authority (WAPDA) which operates a small automated weather station at the base of the glacier [149]. The melt and precipitation data are set at a temporal resolution of 1 day and spatial resolution of 1 deg^2.

5.4.2 Model Setup and Parameters

The melt and precipitation data are read in by MATLAB, projected into the 43N UTM system, and then interpolated onto the SHAKTI mesh. Units are then converted from mm w.e. d^{-1} to m/yr. All water input (ice melt and liquid precipitation) are represented by SHAKTI's englacial input parameter. We continue to maintain the assumption that surface melt is delivered instantaneously to the bed.

This simulation considers only the effect of ice melt calculated from the ERA5 data. The amount and timing of surface ice melt is shown in Figure 5-8. Generally, significant melting seems to start in June every year, although the exact time of year differs; for example, the 2019 and 2020 melt seasons start slightly later than 2017, 2018, and 2021. In addition, 2021 appears to be the biggest melt season, with peak melt in the late summer up to 13 m/yr. The least amount of melt occurs in 2019 when the peak melt is about 8 m/yr.

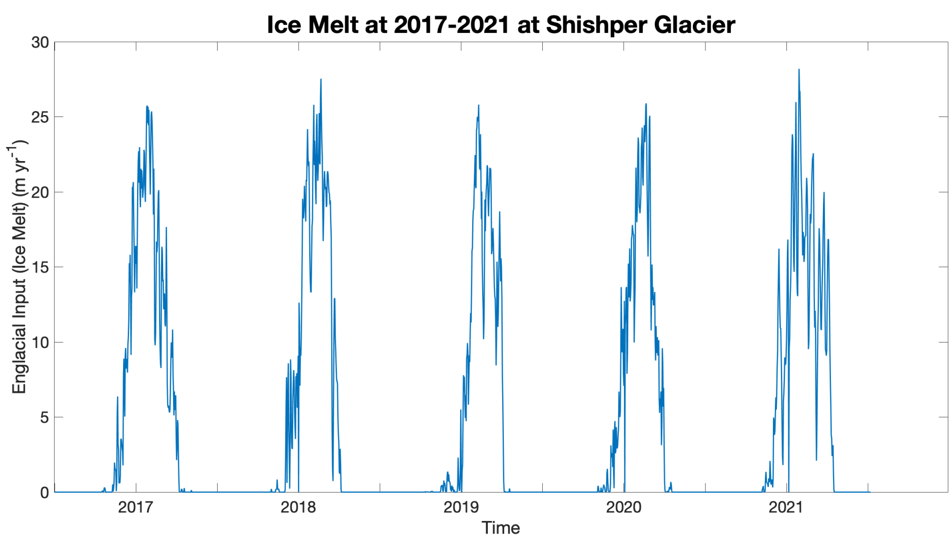

Figure 5-8: *Mean ice melt across the model domain from 2017-2021. Tick marks on the x-axis are placed every six months.*

In Section 5.3.1, we found that adding a uniform, distributed englacial input of 7.0 m/yr to the system required a maximum 30-minute timestep to run successfully. Therefore, we ran the transient simulations with a 30-minute (0.5 hour) constant timestep. Although we observe instabilities in some of the model outputs, none of the numerical instabilities were so great as to require a smaller timestep to keep running. We ran the simulations in ten 6-month chunks on a MacBook Pro 2019, resulting in a simulation runtime of approximately 2 hours per chunk and a total runtime of about 20 hours for all 5 years.

A few of the resulting model outputs (gap height, hydraulic head, basal flux, effective pressure, and Reynolds number), averaged across the entire mesh, are shown below for 2017-2021 (Figures 5-9, 5-10, 5-11, 5-12, and 5-13). An in-depth discussion of the model results is presented in Chapter 6.

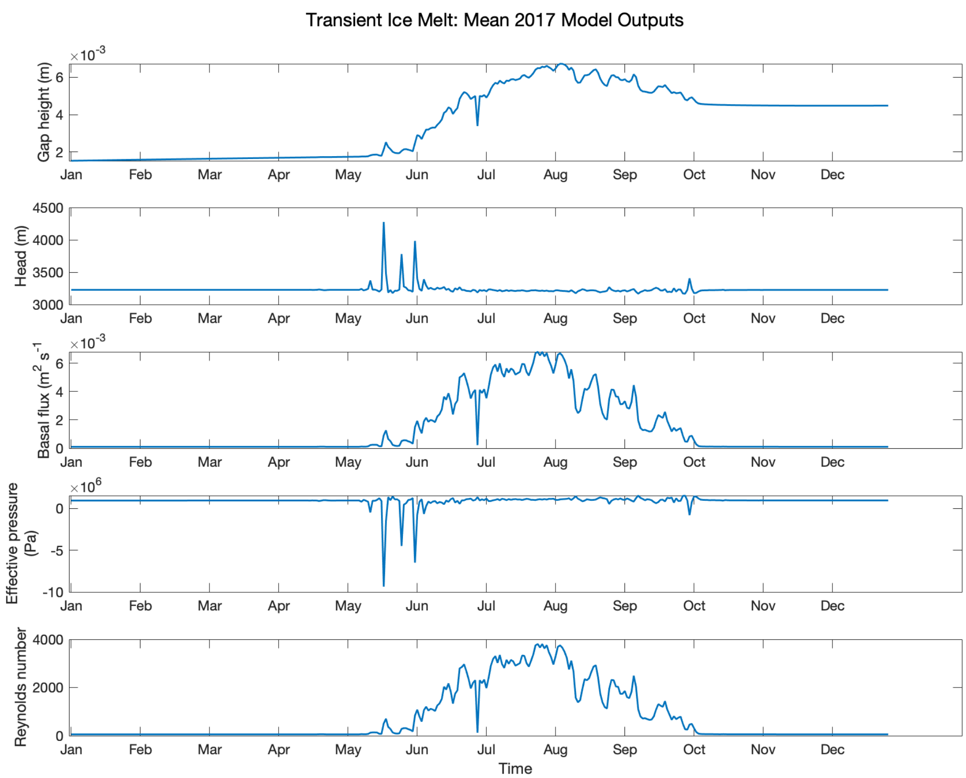

Figure 5-9: *Transient model outputs, averaged across the mesh, for 2017.*

5.5 Chapter Conclusion

In this chapter, we introduced the model we will use, SHAKTI, and the study site for our modeling work, Shishper Glacier in Gilgit-Baltistan, Pakistan. The site was cho-

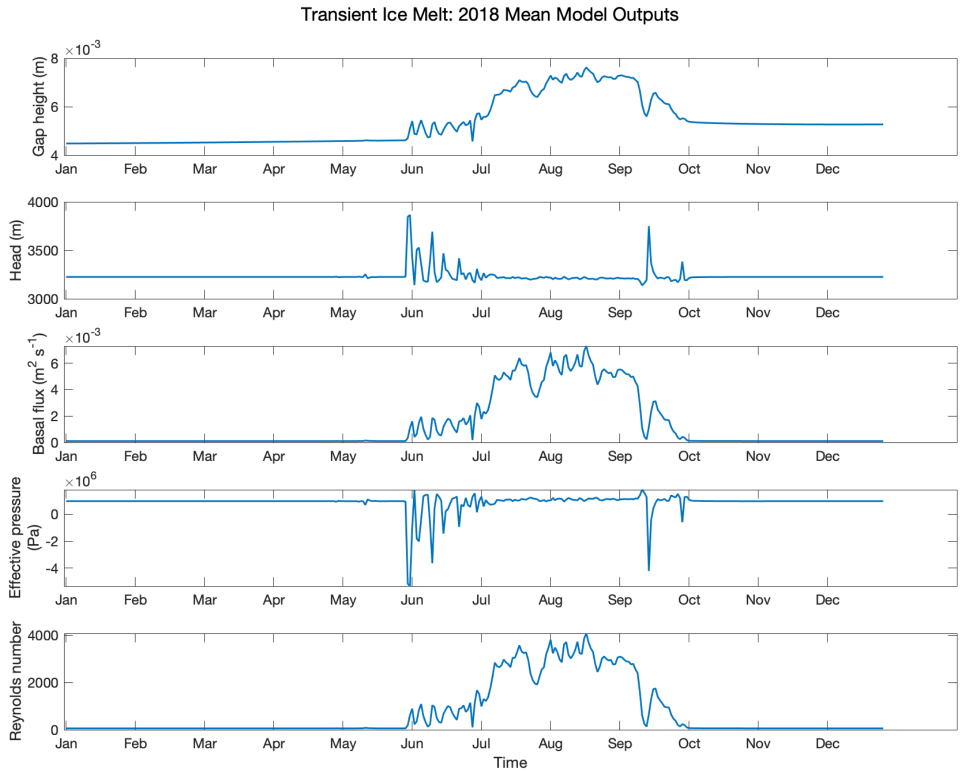

Figure 5-10: *Transient model outputs, averaged across the mesh, for 2018.*

sen due to its varied surge and GLOF behavior in recent years and its close proximity to human settlements and the Karakoram Highway. The several observational studies that have been conducted on it in recent years offer multiple opportunities for model setup and validation. In addition, the formation of Shishper Lake, an impoundment of Muchuwar Glacier's meltwater by Shishper's terminus, is quite similar to several other ice-dammed lakes in the Karakoram. The results of our study may then be useful for those other glacier systems.

Next, we set up the SHAKTI model domain using remotely-sensed data and physics-informed estimates of bed thickness. We simplified the shape of the domain by neglecting Shishper's tributary glaciers. We conducted a winter spin-up, starting from somewhat arbitrary values, to calibrate the model's initial parameters under winter conditions. This showed us the emergence of a central channel running down the main trunk of the glacier, which proves that even when there is no surface melt

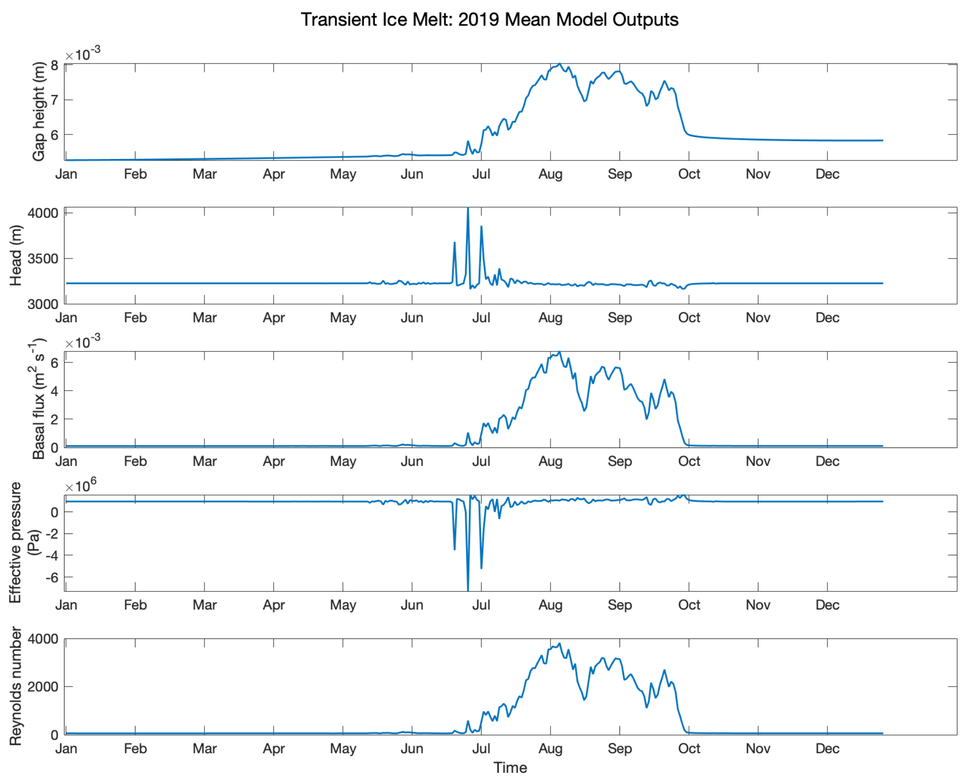

Figure 5-11: *Transient model outputs, averaged across the mesh, for 2019.*

reaching the bed, there is enough basal melt generation occurring for a central channel to be maintained. We then carried out an initial study to investigate the system's response to varied englacial inputs. We find that when we introduce larger melt fluxes to the bed:

1. The channel system reorganizes more drastically, and
2. the simulation requires finer timesteps to converge.

Finally, we set up our transient simulations, which incorporate weather data from 2017-2021. These weather data have been fed through an ice ablation model to provide surface ice melt which can be used as SHAKTI's model input. We then ran the transient simulations and showed an abbreviated form of the results. In the next chapter, we will take a closer look at the results of the 2017-2021 transient simulations

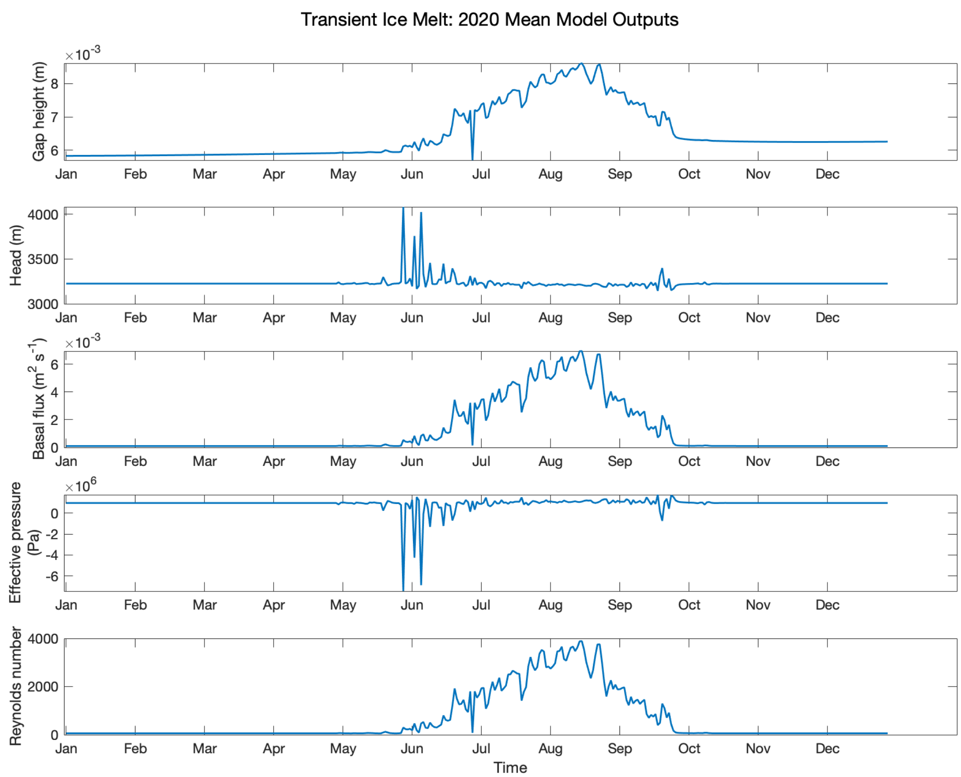

Figure 5-12: *Transient model outputs, averaged across the mesh, for 2020.*

and what they can tell us about surges, GLOFs, and the seasonal evolution of the subglacial hydrological system.

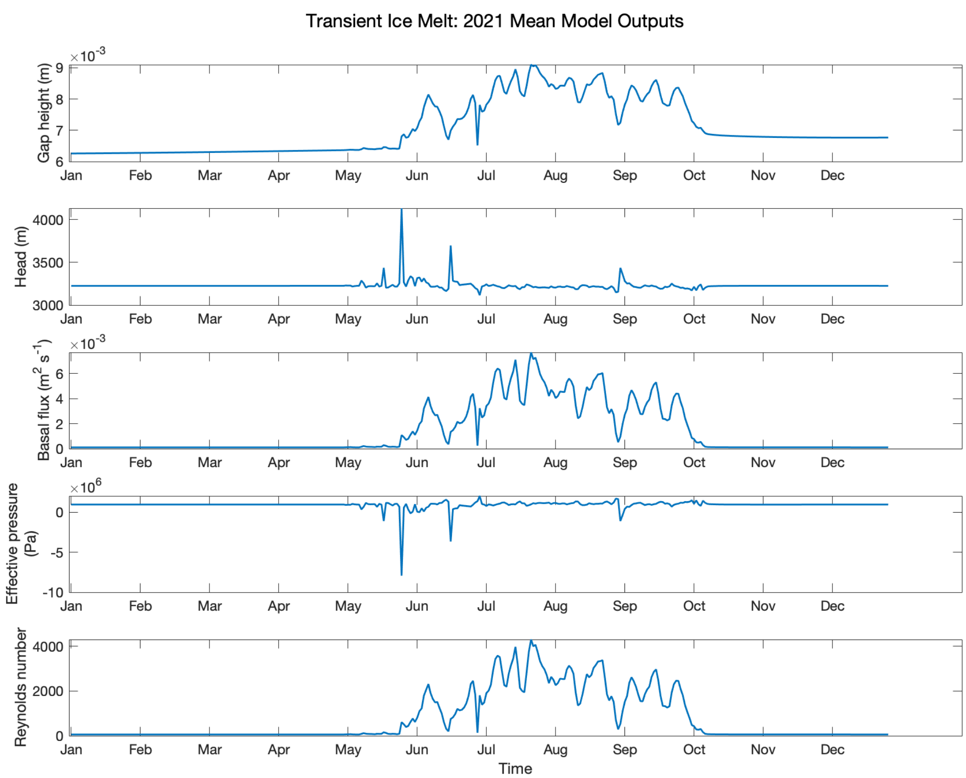

Figure 5-13: *Transient model outputs, averaged across the mesh, for 2021.*

Chapter 6

Shishper Case Study: Results, Discussion, and Conclusions

In Chapter 5, we set up a simplified model domain representing Shishper Glacier in the Karakoram Range in Gilgit-Baltistan, Pakistan. We ran a few preliminary simulations to calibrate the initial model parameters, determine an optimal (i.e., maximum) timestep for transient simulations, and determine whether the model outputs were physically intuitive. In this chapter, we will discuss the results of the transient icemelt simulations and their implications.

6.1 Seasonal Evolution of the Subglacial Drainage System

In Section 5.4, we set up ten six-month transient simulations spanning the years 2017-2021. We chose this timeframe to see if we could observe the conditions before, after, and during the major surge that Shishper underwent between late 2017 and mid 2019. The mean, minimum, and maximum of several output parameters of the model are shown below, separated by year (Figure 6-1).

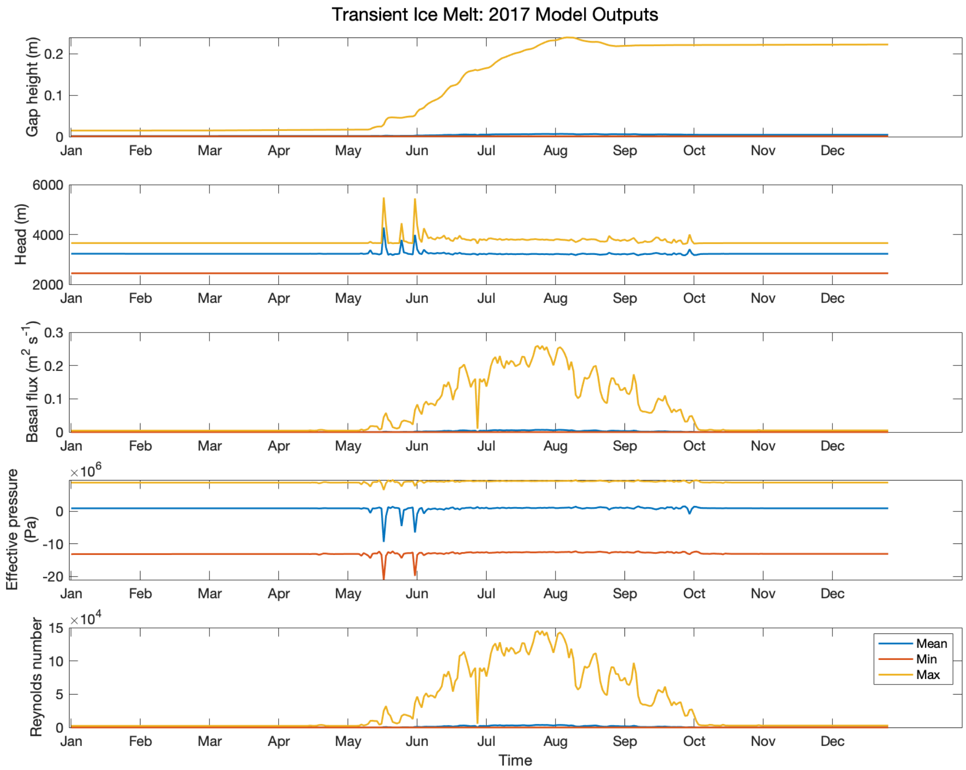

Figure 6-1: *Transient model outputs for 2017.*

6.1.1 Temporal Visualizations of Seasonal Evolution

Gap height: Since the maximum gap height is so much greater than the mean gap height, we can assume that the maximum (shown in yellow) is representative of the main channel, while the mean (blue) represents the rest of the bed. We expect the maximum gap height, which tells us the size of the main Röthlisberger channel down the trunk of the glacier, to be approximately the same size at the beginning of each melt season. After the melt season, we expect the gap height to return to the same winter conditions as the previous winter. However, we can see that the channel grew from 0.02m at the beginning of 2017 to about 0.22 m at the end of 2017, growing to approximately 10x its size. It also grows, albeit much less so, in years 2018-2021. We can conclude that this is simply because the winter spin-up (Figure 5-4) was not run for enough time for the gap height to fully equilibrate. This does not appear to have a drastic effect on model runs; however, for future model runs, we should ensure that

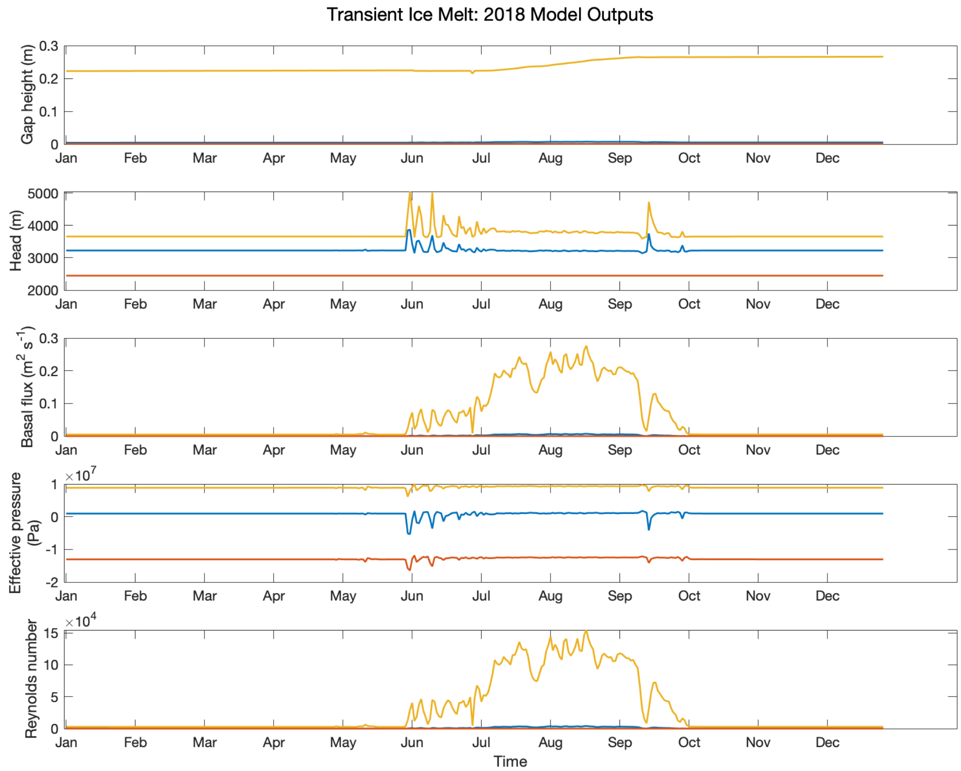

Figure 6-2: *Transient model outputs for 2018.*

the gap height is fully equilibrated at the end of the winter spin-up before running the transient simulations.

Hydraulic head: Hydraulic head h can be defined as:

$$h = \psi + z \tag{6.1}$$

where ψ is water pressure and z is the elevation relative to a specified datum. SHAKTI defines this datum at sea level [146]. Hydraulic head (or just "head") therefore represents both gravitational potential energy as well as potential energy from water pressure gradients. In Figures 6-1 through 6-5, hydraulic head seems to hold constant throughout the simulations with a mean of about 3000m. This is a reasonable value, as the modeled domain spans about 1500 - 3700 m elevation. Every year, there seem to be several spikes during which the head goes above the equilibrium value. In all five years, the majority of these spikes occur at the beginning of the melt

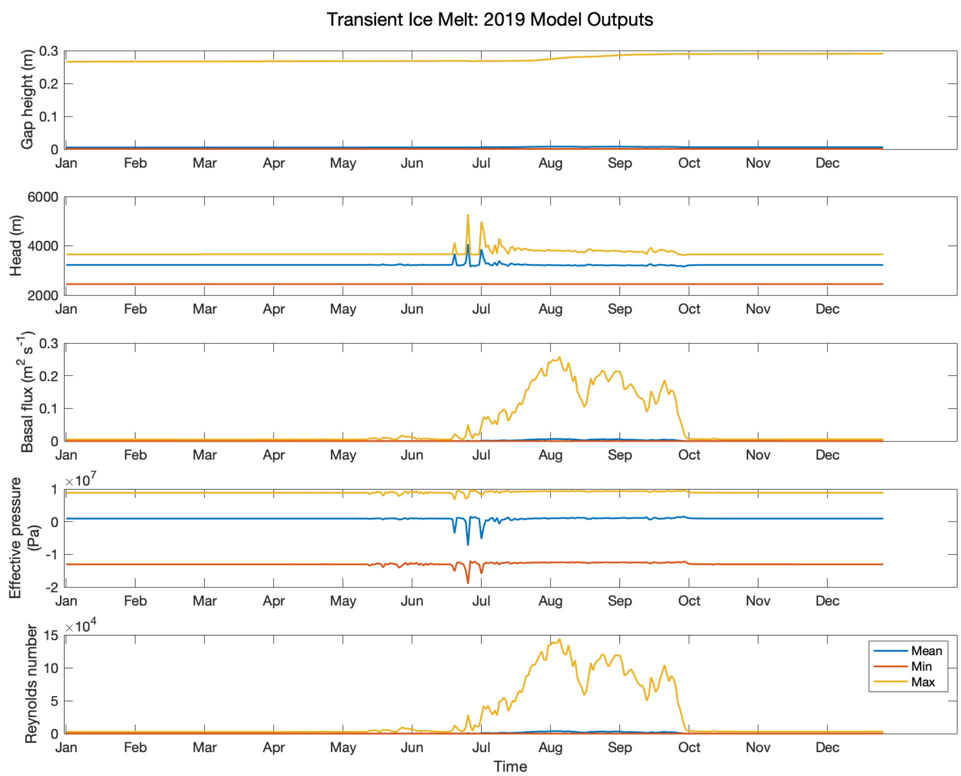

Figure 6-3: *Transient model outputs for 2019.*

season when basal flux starting to rise, and all of the spikes occur during the melt season - i.e., periods when the basal flux is greater than 0. The big spikes of higher hydraulic head at the beginning of the melt season (May-July, depending on the year) also indicate that the system is still distributed (inefficient); the rise in head shows a buildup of water pressure due to the system's inability to transport growing fluxes, and the rapid fall in head back to the equilibrium value shows that the system resolved this pressure by creating or growing a new, efficient channel.

We also observe smaller spikes in head towards the middle to end of the melt season (August-September). According to [34] and [50], multiple small drainages have been observed to occur from subglacial drainage networks as the subglacial system contracts at the end of the melt season. This is probably what we are seeing here. This contraction, or closing of the system, happens as basal flux falls, allowing melt opening to fall and creep closure to dominate. As the system closes, it becomes less

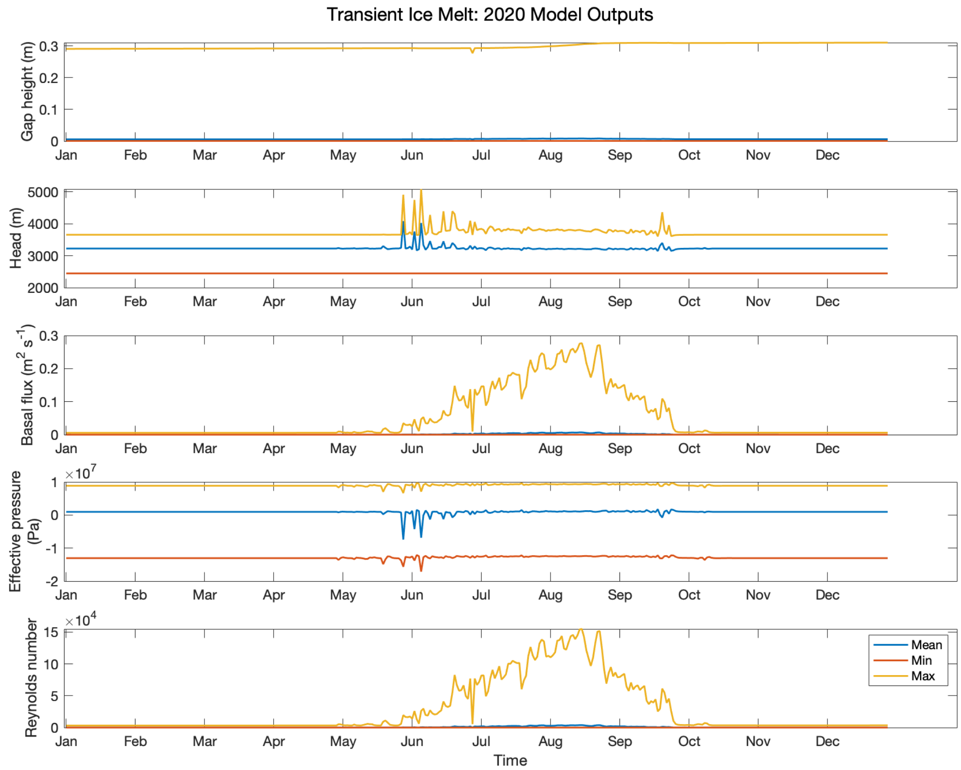

Figure 6-4: *Transient model outputs for 2020.*

efficient at transporting fluxes, resulting in these small spikes in pressure. The spikes are smaller than the ones at the beginning of the melt season because the system has not had much time to close yet, so the system is still more efficient than it would be at the beginning of spring.

These spikes in water pressure, which only occur when the system is inefficient and experiencing high fluxes, may also lead to drainages at the outlet of the glacier. However, it is difficult to tell from this model whether this is true. This may be a subject for future work.

Basal Flux: Similar to gap height, the maximum basal flux is much higher than the mean basal flux across the glacier, meaning we can take the maximum (yellow) line as a representation of what is happening in the main channel/channel system. Similar to what we saw in Figure 5-8, the melt season starts at a slightly different time each year due to differences in weather patterns. Since the model assumes

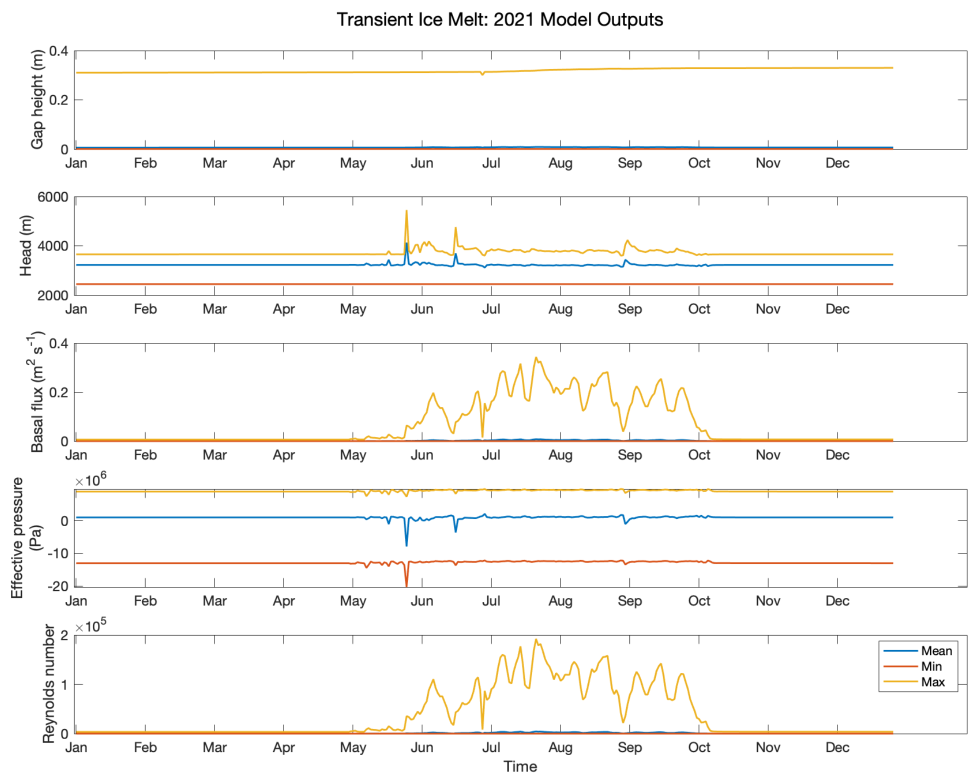

Figure 6-5: *Transient model outputs for 2021.*

instantaneous delivery of surface meltwater to the bed and no englacial storage, we can assume that the melt season begins exactly when basal flux begins to rise. In 2017, the melt season begins much earlier, in mid-May (Figure 6-1); in 2018, 2020, and 2021 (Figures 6-2, 6-4, 6-5) the melt season starts in late May to early June, and in 2019, the melt season starts quite late, in late June/early July (Figure 6-3. However, all melt seasons end at nearly the same time, at the end of September. The 2017 melt season was by far the longest one of the five years (approximately 150 days), and 2019's was the shortest (approximately 100 days).

Effective pressure: The spikes in hydraulic head correspond with dips in effective pressure, during which time effective pressure becomes negative. Recalling the definition of effective pressure from Section 4.1.2, we indeed expect the effective pressure to drop when water pressure rises. When effective pressure is negative, it means that water pressure exceeds the ice overburden pressure. Therefore, we expect to see glacial uplift, a phenomenon which has been observed by [43] and [90]. We will discuss these changes in effective pressure in 6.1.2.

Reynolds number: Reynolds number Re is defined as

$$Re = \frac{uL}{\nu} = \frac{|q|}{\nu} \tag{6.2}$$

where u is the velocity of the fluid (water), L is the characteristic length scale (defined as gap height b) of the channel, q is depth-integrated basal flux, and ν is the kinematic viscosity of the fluid. A high Reynolds number indicates more turbulent flow, while a low Reynolds number indicates laminar flow. In our case, ν of water stays constant, so the variables of interest are the velocity and the characteristic length scale. In all five simulated years, the maximum of Reynolds number very closely follows the trend of maximum basal flux. This makes a great deal of physical sense, as channels with greater flux likely also travel at faster velocities, leading to higher turbulence. In addition, due to melt opening (turbulent dissipation), the channel width also scales with flux q, increasing L which also increases Re.

Response Time to Changes in Melt Delivery

Understanding the system response time to melt inputs can provide an additional insight into how the system responds to triggers. For example, Figure 6-6a shows a time series of englacial input (surface melt input to the bed) and basal flux during the year of 2017. Both fields are averaged over the entire domain, so the basal flux series represents the cumulative system response to an overall englacial input across the mesh. The basal flux, shown in red consistently and closely lags melt input (blue). A similar trend is seen for years 2018-2021. Cross-correlation of the peaks reveals that the characteristic lag time for the entire system is about 1 day (Figure 6-7a). This means that it only takes about 1 day for the mean basal flux across the mesh to respond to changes in the mean englacial input.

Comparing the characteristic lag times across different model outputs, areas of the domain, and times of year can give us a better understanding of the heterogeneity of

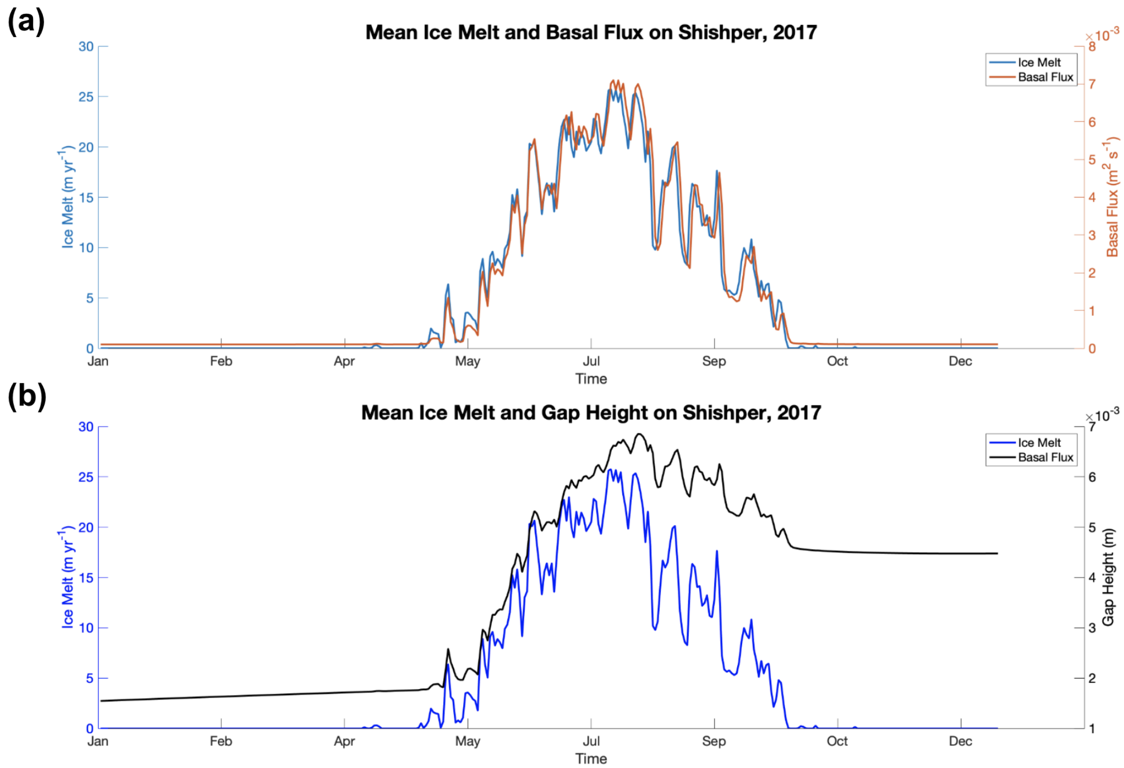

Figure 6-6: *(a) Mean surface melt input and basal flux during 2017. (b) Mean surface melt input and gap height during 2017.*

the system response. For example, all of the cross-correlation plots between ice melt and basal flux for 2018-2021 are very symmetrical and show a neat peak (for example, Figure 6-7a). However, the characteristic lag between ice melt and gap height show interesting asymmetries, even though the lag time is also about 1 day. For example, Figure 6-6b shows the lag between ice melt and gap height. It is less clear-cut than the relationship between ice melt and basal flux (Figure 6-7a) because the gap height has a nonlinear response to the system input (recall Equation 4.10).

The sampling rate of the model outputs is one data point per day; therefore, we are not able to obtain a more precise lag time. It is likely that the lag time is less than a day; however, a better resolution of output data is necessary to narrow down a more precise lag time.

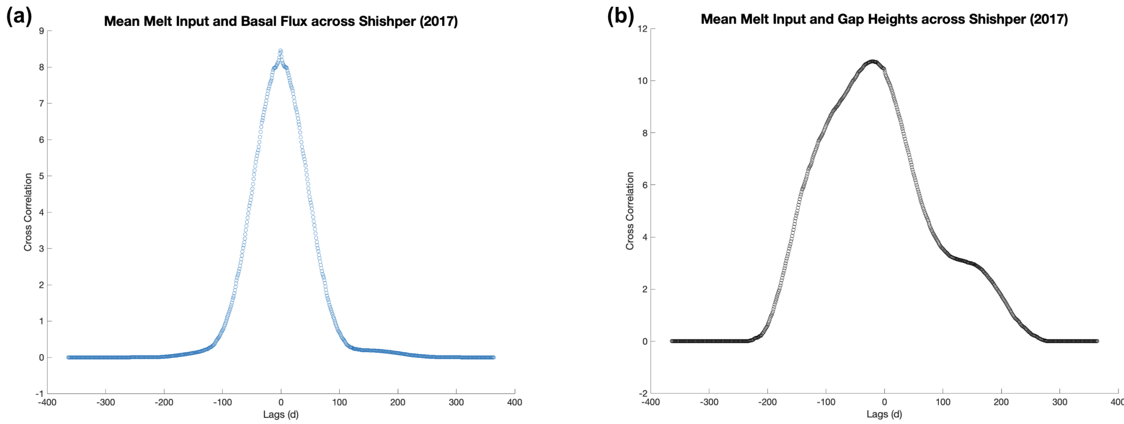

Figure 6-7: *(a) Cross correlation of mean ice melt input and mean basal flux across the mesh shows a peak cross correlation at lag -1, meaning that basal flux lags ice melt by 1 day. (b) Cross correlation of mean ice melt and mean gap height across the mesh show a peak correlation at lag -1. Since the sampling rate is once per day, we do not have a more precise lag time.*

6.1.2 Spatial Visualizations of the Subglacial Drainage System

From the time-series outputs of the transient model (Figures 6-1, 6-2, 6-3, 6-4, and 6-5), we were able to glean information about how the efficiency of Shishper's drainage system evolved over time. However, since these plots are averaged over the entire mesh, it is difficult to investigate where changes are occurring in the domain. In this section, we will visualize how these changes occur spatially over Shishper Glacier.

Overview of Seasonal Changes

Spatial plots of the transient ice melt model solutions demonstrate the evolution of the subglacial drainage system we looked at in section 6.1.1 (complete videos of this evolution by day can be found in the Supplementary Materials). Figure 6-8 shows the configuration of the drainage system throughout 2017. The first plot (Figure 6-8a) shows the system on January 1, 2017, in a mostly closed state. In this case, drainage channel in the middle to lower part of the domain is observed, although it appears thin without a lot of flux. Figure 6-8b shows the system on August 18 at the height of the melt season on a day during which the system is handling the most amount of water. At this peak, the channel has extended upwards toward the higher-elevation parts of the domain, splitting in two around 4030 km north in an arborescent pattern (i.e., tree-like, branching). This type of arborescent drainage pattern was proposed by Röthlisberger [130] and is characteristic of an efficient (channelized) drainage system. The presence of this efficient system is expected for a steep glacier like Shishper, where head potential gradients are steep (recall section 4.1.5). Figure 6-8c shows the system late in the melt season on September 12; in comparison with (b), several branches of the upper part of the system have disappeared or thinned. By October 18 (Figure 6-8d) the upper part of the system has completely shut down, leaving only the lower part of the channel still open, and with significantly less flow going through.

The lower channel which traverses the mid- to lower trunk clearly persists through every simulated winter in 2017-2021, and can be seen in both images of the "closed" state (Figure 6-8a and d). Recalling Figure 5-5, we know that this channel appears during the winter spin-up, during which time the only water at the ice-bed interface comes from pressure-induced melting and geothermal heat flux (frictional heating is 0 in all of our simulations thus far because ice velocity is set to 0). Because Shishper is a temperate glacier, parts of the basal interface are usually able to be maintained at the pressure melting point for most of the year [67]. Therefore, there is always a consistent stream of water, although small, that keeps the main channel open. It also makes sense that the channel in the lower part of the domain stays open; all of the meltwater across the glacier bed is directed downhill due to gravity. [26] show that surface melt elevations move from 6400m in peak summer to 3500 m at the end of winter (no surface melt is observed in December, January, or February) meaning that the bottom part of the glacier will always receive more melt than the top.

Channel Formation During the Early Melt Season

In Section 6.1.1, we observed that large spikes in hydraulic head occur at the beginning of every melt season in our simulations. We postulated that new channels are created during these spikes in pressure. To investigate the effects of these channels, we will take a look at the spike in pressure that occurred during early June 2017 (see Figure 5-9). Figure 6-9 shows the state of the system during that time. On June 3 (6-9a),

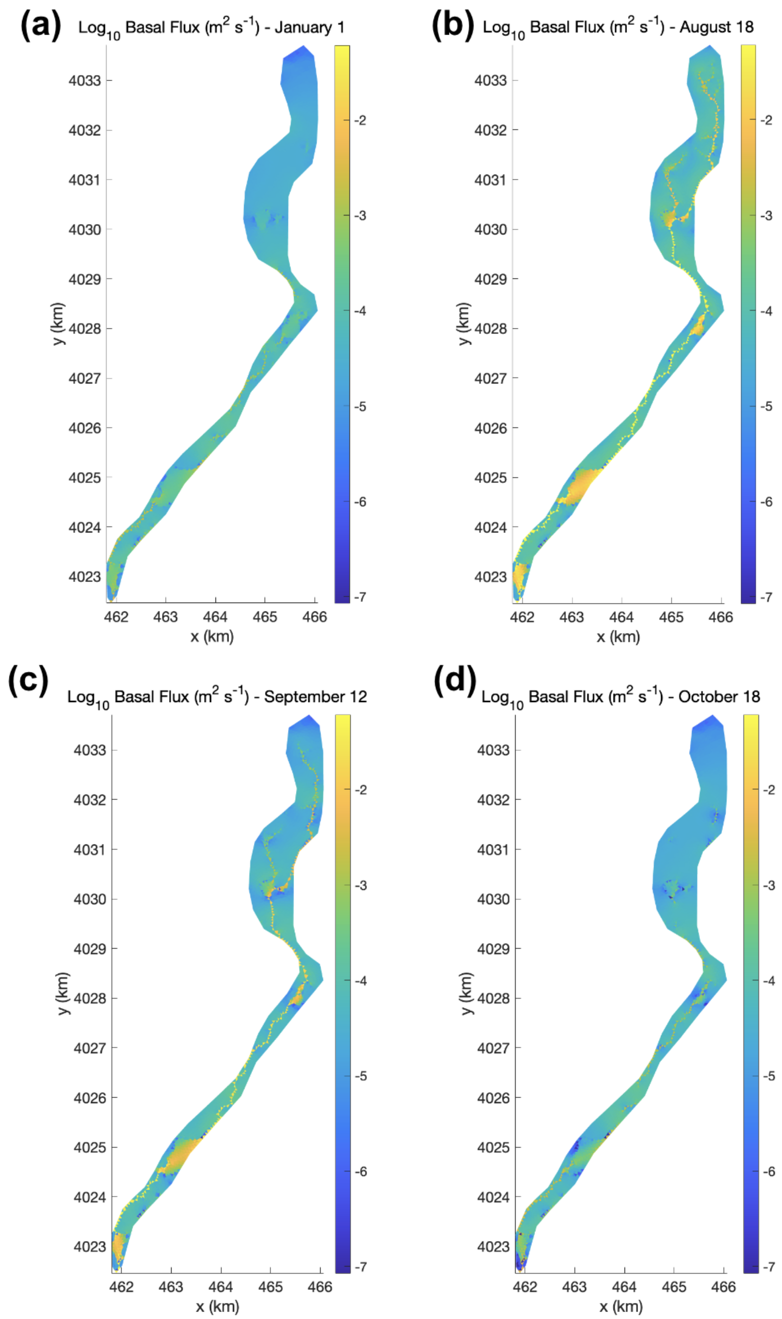

Figure 6-8: *Basal flux (shown in log scale to emphasize smaller channels) across the mesh at various points in 2017; (a) January 1, (b) August 18, (c) September 12, and (d) October 18.*

the system shows relatively channelized flux that is concentrated in the main drainage channel down the middle to lower part of the domain. By June 5th, the system has become far more distributed, showing raised fluxes across the domain (yellow/green).

Between 4029 and 4030 km north, we see a particularly thick area where there is a slightly concentrated and heavier flow (indicated with a red arrow). Between June 5th and June 10th, the flow has narrowed to a more concentrated area. There is also less basal flux to the right and left sides of this increasingly channelized flow. By June 20th, this area of increased flow has turned into a new channel.

We interpret this sequence of events as follows: During the 17 days between June 3rd and June 20th, an area of high flux and high pressure develops, forming a distributed, sheetlike flow. The wide sheet of high flux quickly narrows as it becomes an efficient channel, which then connects with the previously existing Röthlisberger channel below. The formation of this channel extension results in the efficient removal of flux from the surrounding area, decreasing the local water pressure over the subglacial interface and concentrating most water flow to a single, narrow channel.

The behavior of this modeled system clearly corroborates the understanding in the scientific literature that there is a transition from a distributed to channelized drainage system during the early melt season (for example, [135, 34, 67]). Furthermore, speed-ups in surface displacement further up the trunk of Shishper were observed by [26] during the early melt season (May to June) between 2013-2016, indicating that there is increased basal water pressure at the northern part of the domain during this time. This agrees with our model results: near the terminus, the system remains perennially channelized, while the upper part sees an inefficient, distributed system during the early melt season. We only expect to see high water pressures at the upper parts and not near the terminus. This is an encouraging result because it shows us that our model is well-suited to provide insights into surface displacement and surge-type behavior.

Emergence of Uplift During the Early Melt Season

During these spikes in head, we also observe a decrease in effective pressure N (as seen in Figure 6-1). As head increases, the effective pressure decreases, in several areas becoming negative. Recalling the definition of effective pressure from 4.1.2, $N = p_i - p_w$, we can understand that N becomes negative when p_w, the water pressure,

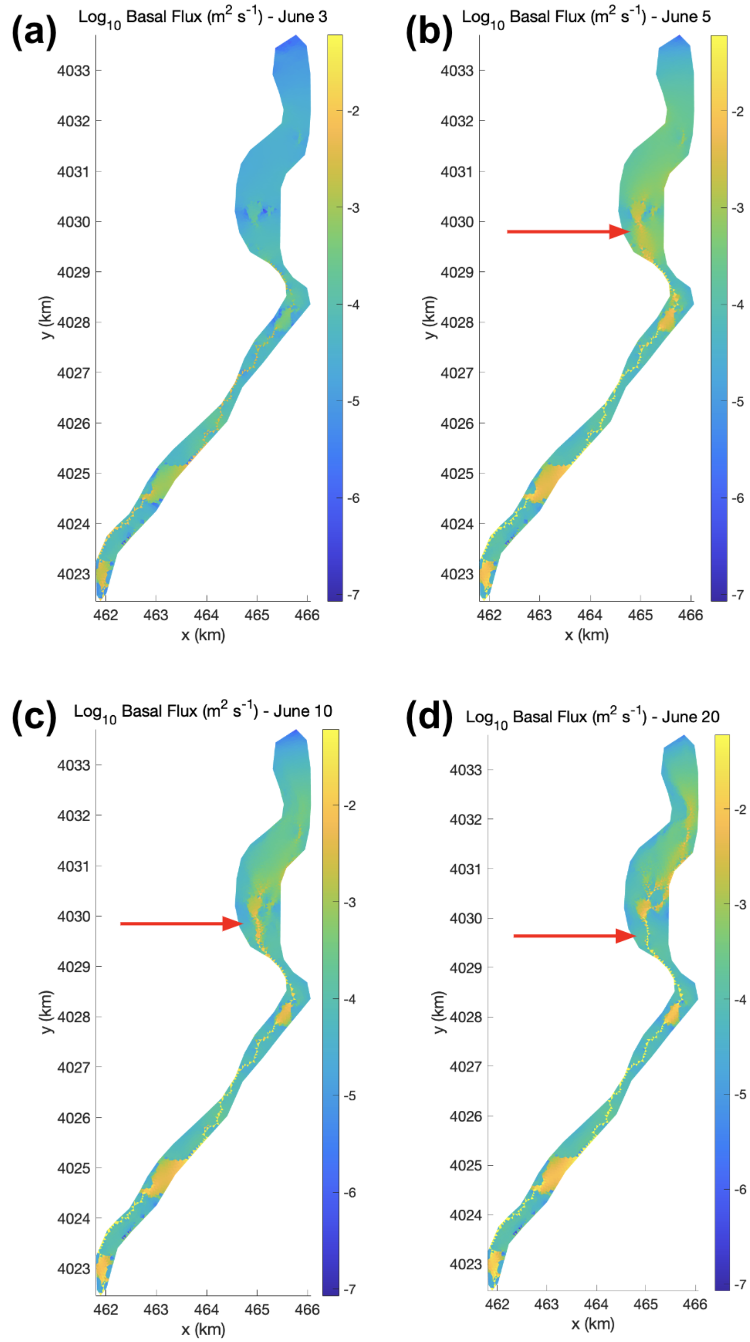

Figure 6-9: *Basal flux (log scale) on (a) June 3, (b) June 5, (c) June 10, and (d) June 20, 2017 shows the creation of a new channel in response to increased englacial input.*

exceeds p_i, the overburden pressure of the ice. Depending on the mechanical strength of the ice and the underlying till or bedrock, this negative effective pressure can either lead to ice uplift or hydrofracturing of the underlying material [146].

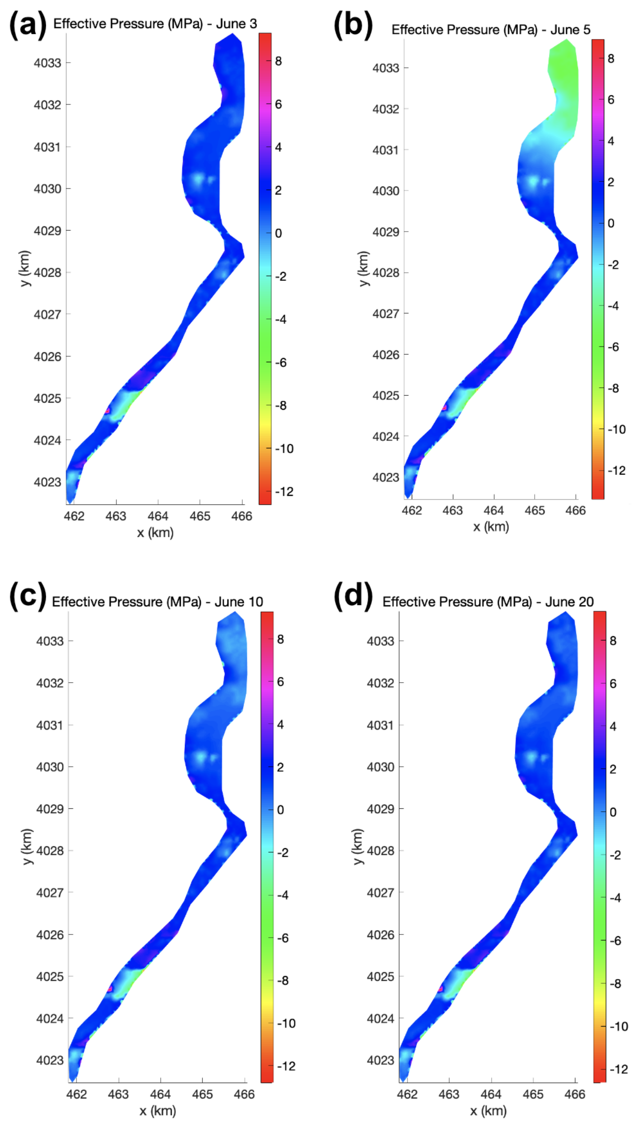

Figure 6-10: *Effective pressure at Shishper on (a) June 3, (b) June 5, (c) June 10, and (d) June 15. Note the negative effective pressure observed at the upper part of the domain on June 5.*

Figure 6-10 shows the effective pressure over the domain during the early-melt season spike (cf. 6-9). Figure 6-10a, the "base" state of effective pressure on June 3, shows that effective pressure is approximately 2 MPa across the mesh, apart from a few high-pressure areas that correspond with cavities in the bedrock. The spike, or drop in effective pressure can be clearly observed in 6-10b (June 5), where a large area of negative N can be seen in green at the northern part of the domain. Recalling our above discussion of 6-9, a channel formed in the upper part of the domain between

June 5 and June 20, allowing that section of the system to become more efficient. On June 10, this area of uplift at the northern part of the glacier has lessened, and by June 20 the entire section has almost completely returned to the original state of effective pressure, around 2 MPa across the mesh.

The presence of uplift has a few interesting implications. Uplift has been observed to immediately follow supraglacial lake drainage [35, 90, 43]. Although the exact phenomenon of supraglacial lake drainage is not modeled in this work, it represents a rapid delivery of surface melt to the bed, which we are modeling in this simulation (in supraglacial lake drainage, the delivery of meltwater to the ice-bed interface is concentrated at a single point; in our simulations, it is distributed across the mesh). The vertical movement of ice during uplift can introduce mechanical stresses into the glacier, which has also been observed to cause fracture, a phenomenon also known as hydraulic jacking [35]. When this occurs close to the terminus, it could result in calving of the glacier. This has implications for both the ice dynamics and for flooding; a rapid loss of ice mass near the terminus may decrease the buttressing forces that restrict the glacier's downward flow. In addition, a rapid transfer of volume into any glacial lake below the glacier could lead to dam overtopping and GLOF [111]. The feedback processes between calving, surging, and GLOF are very complex and should be investigated further in future modeling work (see Section 6.2.3).

In our case, we see negative N (likely uplift) at an area that is much higher up in the glacier. [90] found that hydraulic transmissivity K of the subglacial hydrological system could be inferred by the uplift relaxation time. They observed that early-melt season drainage events were followed by slower uplift relaxation than late-melt season drainage events, due to a lower relaxation time. Figure 6-11 shows the hydraulic transmissivity across the system after the "spike" observed around June 5th (early melt season) and around September 1 (late melt season). The transmissivity plots show that the system is much more efficient in September than in June, allowing high water pressures to resolve more quickly and for effective pressures (uplift) to relax more quickly. These results give support to previous studies that suggested that surface uplift relaxation times scale with transmissivity of the subglacial drainage

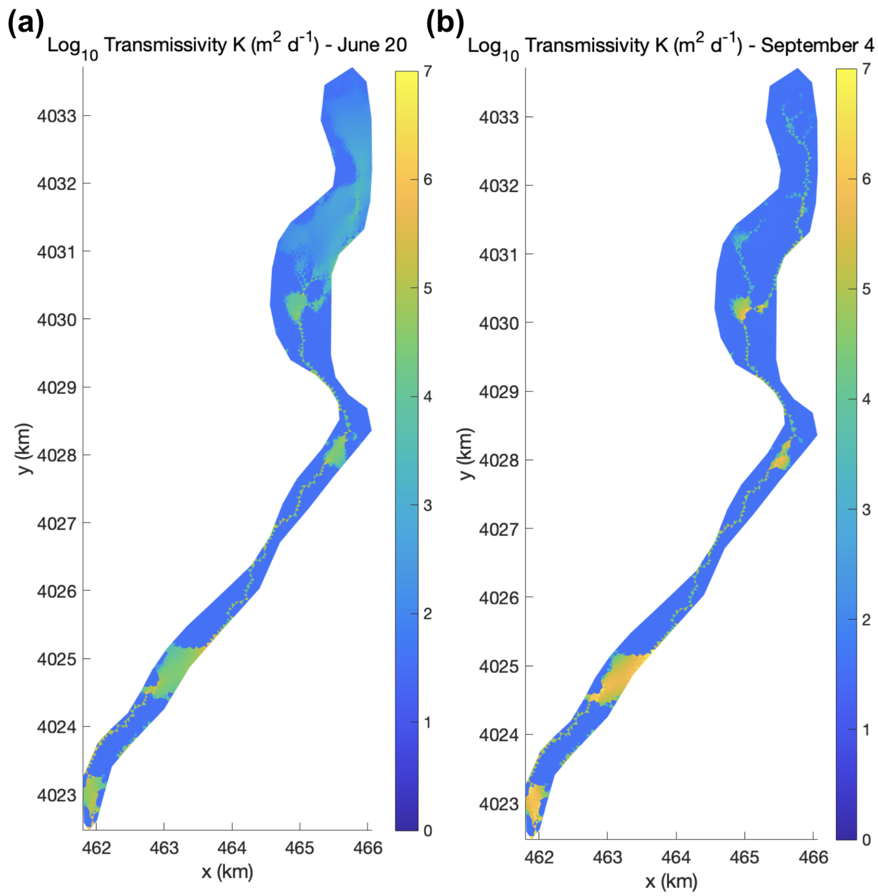

Figure 6-11: *Transmissivity K after (a) an early melt season spike and (b) a late melt season spike.*

system [90, 45]. However, SHAKTI is solely a subglacial hydrological model, and these simulations are not coupled to ice dynamics, so we cannot make a definitive claim that a more strongly negative N leads to greater uplift. A future study on this phenomenon is suggested in Section 6.2.4.

System Closing During Late Melt Season

As the melt season draws to a close, we observe the closing of channels during which the subglacial drainage system returns to its closed winter state. Recalling Figure 6-8, we can see that channels begin to close around mid-September and the system almost completely returns to its winter state by mid to late October. This corresponds with lower surface melt inputs, which decrease the melt opening term (Equation 4.10).

Creep closure becomes the dominant mechanism controlling channel size (gap height), which leads to a quick decrease in gap height over time. The channels at the northern (upper-elevation) section of the glacier disappear first, due to lower cumulative basal flux and lower surface temperatures compared to the lower part of the glacier. As mentioned earlier, the lower channel remains open throughout the winter. Consistent yearly observations of small winter drainages at Shishper's base in December and January provide evidence that water continues to flow through the subglacial drainage system throughout the winter, supporting our claim that the main channel persists through the winter [149].

6.1.3 Surging and GLOFs between 2017-2021

As discussed in Chapter 3 and 4, the field of glaciology has drawn a clear link between subglacial hydrology and the phenomena of glacial surges and glacial lake outburst floods (GLOFs). In particular, at our modeled Shishper Glacier, both GLOFs and surges occur regularly. Now that we have established that our simulated system behaves physically, we can investigate the link between subglacial hydrology and surging/GLOFs.

Shishper underwent a major surge between 2017-2019, during which time the terminus advanced approximately 1.5 km [26]. [26] used satellite imagery to perform velocity tracking to assess when and how surges have occurred on Shishper in the past few decades. They found that prior to 2017, the glacier also surged between 2000-2001 and in 1973. As noted earlier, they also found significant seasonal speedups in the middle of the glacier trunk (not at the terminus) during the early melt season [26]. A full "surge" is then defined as the movement of the entire glacier rather than just a section further away from the terminus, which always remains efficient/channelized. Figure 6-12 shows a timeline of recent surge behavior documented by [26].

Figure 6-13 shows model outputs overlaid with approximate surge velocity phases as described by [26]. Spaces between the highlighted areas show times when velocity data was not available due to cloud cover. According to [26], their observations show a pre-surge acceleration between November 2017 to February 2018 (highlighted

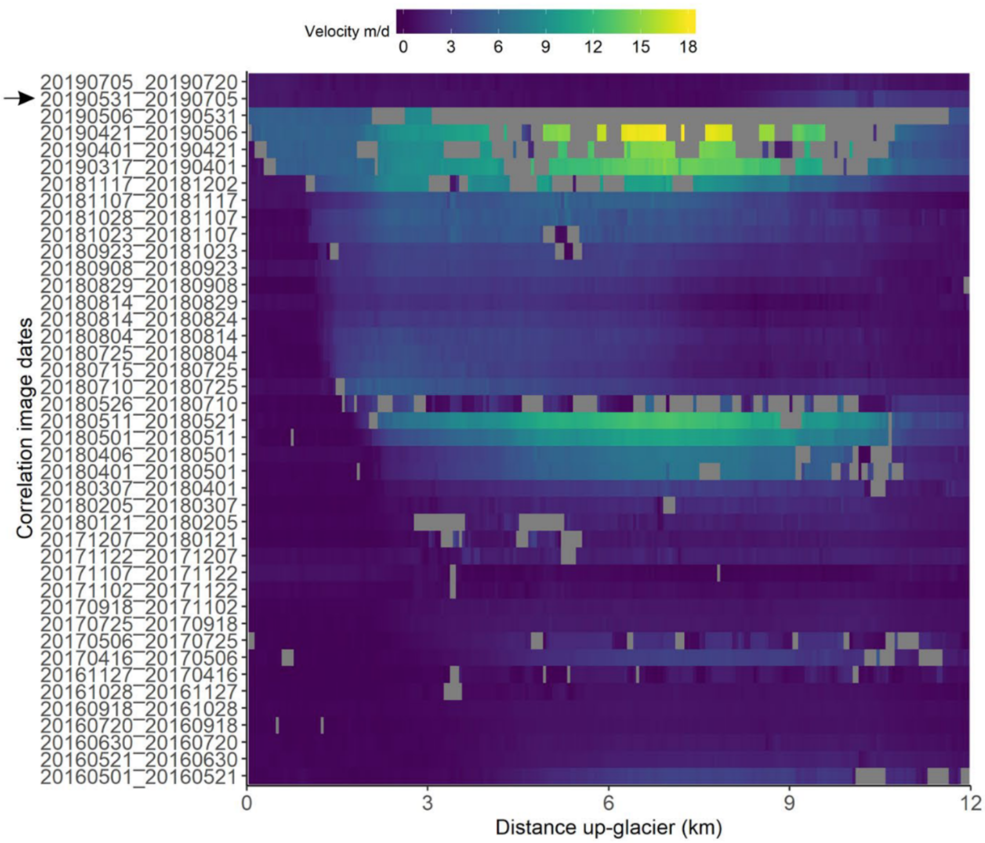

Figure 6-12: *Figure from [26] showing variation in surface displacement across Shishper. The black arrow at the top left indicates a GLOF in June 2019.*

in yellow). When this period begins, it is well after the melt season (basal flux is approximately 0 over the domain) and there are no major spikes in hydraulic head. There is no indication given by the model outputs that would suggest an explanation as to why the pre-surge acceleration started at this time (e.g., an increase in pressure leading to bed lubrication). An active phase of the surge is seen by [26] between April 6, 2018 and May 11, 2018 (highlighted in orange), during which speeds reach approximately 6 ± 0.1 m/d. Similarly, since this happens during the winter season and there is no basal flux or rise in hydraulic head at this time, our simulation provides no hydrological reason as to why the glacier velocity increased. At the beginning of June 2018, [26] describe a rapid but brief acceleration to 18 ± 0.5 m/d (highlighted in red).

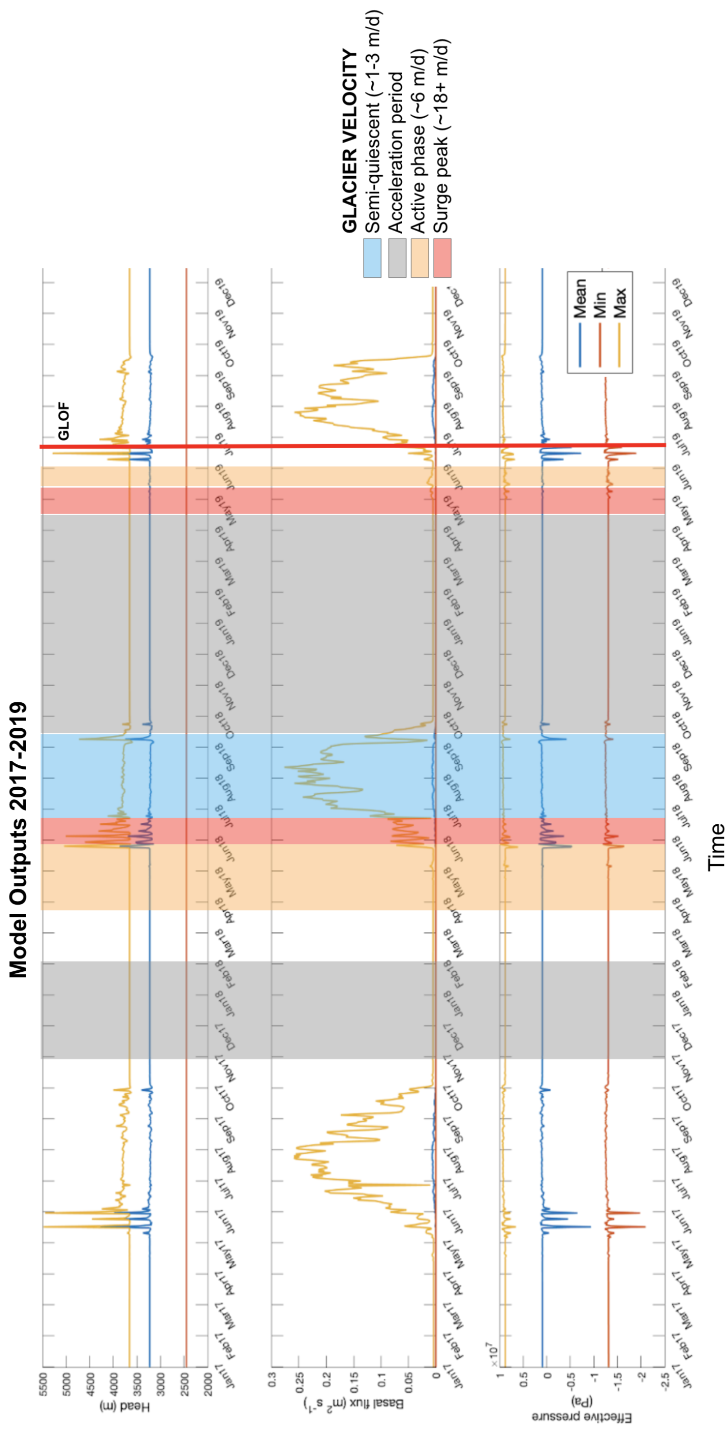

Figure 6-13: *Surge phases analyzed via satellite imagery by [26] overlaid on model outputs of head (m), basal flux (m^2/s) and effective pressure (Pa). The bright red line indicates a GLOF that occurred on June 22-23, 2019.*

This does coincide with a series of "spikes" in head at the beginning of the 2018 melt season, during which time hydraulic head (water pressure) spikes dramatically as it does at the beginning of every melt season. The end of this surge peak is marked by a return of the system to a low-head state as the drainage system becomes efficient and channelized. The surge then enters a very slow period during which velocity is only slightly higher than normal summer velocities (Figure 6-12 in blue). This continues until mid-September 2018, when the glacier begins to accelerate again to speeds of 2.5 m/d. By October 2018, speeds increase to about 3.5 m/d (highlighted in gray). This period follows a slight peak in head at the beginning of September 2018. Finally, another surge peak, again moving at 18 $\pm$ 0.3 m/d, follows on April 21, 2019, lasting until May 6 (in red) [26]. A small GLOF, which damaged the Karakoram Highway, follows on June 22-23, 2019 (denoted by a bright red line).

Most of the changes in glacier velocity observed by [26] occurred without a hydrological trigger that was obvious from our model outputs (i.e., surface ice melt alone). However, the beginning of the surge peak in June 2018 coincided with a large spike in hydraulic head during the beginning of the 2018 melt season, and the start of the long "yellow" period in September 2018 coincided with a spike in head at the end of the melt season. This indicates increased bed lubrication which may be able to explain a speedup of surge velocities. It has long been suggested that subglacial hydrology plays a critical role in surge behavior, as [80] first proposed in 1987. However, the lack of a subglacial hydrological trigger for the 2017 incipient surge motion and 2019 surge peak – at least when considering surface ice melt alone – suggests that rises in water pressure at the ice-bed interface may not provide a standalone explanation for surge behavior.

It is very possible that the spikes in head that coincided with surge speed-ups could have played an important role in those speed-ups, however. If a build-up of potential energy (i.e., through mass or enthalpy accumulation) or dynamical thinning [100] occurred such that the system was at an elevated state of surge "risk", a hydrological trigger could set off the surge motion. Either way, our results indicate that subglacial hydrology likely plays more of a role in *stopping* surges than starting them. The first active phase/surge peak in Figure 6-13 was terminated when the system switched to a channelized system and reached a low-pressure state. The second active phase also ended following a lake drainage in June 2019.

It is possible that if we couple the model with sliding velocity and thus include frictional heating, there would be more water in the system that could change how quickly the drainage configuration deals with influxes of water. In addition, our current model does not include liquid precipitation; the monsoon season which occurs from April to July may possibly play a limited role in some speedups such as the one that occurred in April 2018. A final unknown is englacial storage and release of water, which we do not model in this simulation. A description of future work that may shed further light on surge behavior is described in Section 6.2.5.

6.1.4 Model Limitations

While models can provide crucial insight into the evolution of subglacial processes, they have critical limitations. In this work, there is a lack of observational data both to calibrate the model parameters and to validate the results. Much of the data used to calibrate the model are modeled themselves, rather than from direct observation: for example, the glacier thickness is inferred from surface topography, and the ice melt data is modeled from interpolated weather data which is then channeled into an ice ablation model. To validate results pertaining to the evolution of the subglacial system, we can compare the intuition the model gives us to the results from observational studies of subglacial hydrology. However, there are many parameters of the model that remain unknown that can strongly affect the response of the system to inputs: for example, the creep parameter, englacial storage times, geothermal heat flux, sliding parameters, and most of all, the composition of the underlying bedrock.

The lack of knowledge of the underlying bedrock is a crucial limiting factor in these experiments. Parametrizing the behavior of subglacial sediments and their role in subglacial hydrology is highly dependent on understanding the composition of the till. SHAKTI assumes that the underlying material is solid bedrock; however, it is largely unknown whether this is actually true in the Karakoram. Our model assumes that all channels at the ice-bed interface are cut into the ice rather than into the underlying bedrock or till (N-channels); hydrological flow through porous sediments is also neglected. This assumption likely causes the model's behavior to deviate significantly from the physical system; however, it is difficult to tell what these errors are and how to correct for them.

The model also does not account for any englacial processes. As mentioned multiple times in our discussion, our implementation of SHAKTI assumes that surface melt instantaneously reaches the bed. As discussed in Section 4.1.1, a complex range of englacial processes affect the delivery of meltwater to the bed. However, SHAKTI takes a crude approach to parametrizing englacial water storage because we do not have enough observations of englacial storage to better understand the role in englacial channels, firn aquifers, and snow to ice compaction. The configuration of these englacial features likely affect the release and arrival of flux to the bed, which are not represented in this work.

Another major limitation of the current model formulation is that we have consistently prescribed the velocity of the glacier to be 0. From numerous observations at Shishper, it is quite obvious that this is an unphysical assumption [26]. The model assumes that surface velocity is equal to basal velocity, which is an appropriate, but not perfect assumption for fast-moving glaciers such as Shishper [147]. The introduction of surface velocity affects the subglacial drainage system significantly. For example, it introduces basal drag, which creates frictional heating, thus introducing much more melt into the system. The movement of the glacier also introduces increased cavitation which can allow water to collect in large subglacial pools, and may destroy existing channels, hindering the system's ability to evolve to a fully efficient state.

6.2 Future Work

The results of this work so far have been quite encouraging. They agree with previous scientific intuition and observations of subglacial hydrology. They show that this work can serve as a good starting point for further, more sophisticated modeling work on subglacial hydrology in the Karakoram range. The results of our study lead to several immediate and exciting next steps to improve our understanding of Shishper's subglacial system. We will talk about some of these next steps here.

6.2.1 General Improvements to Model Accuracy and Precision

There is much work needed to improve the rigor of our setup and the accuracy of the results. There have been a few flaws in our methods which can be readily corrected in future work. A few quick relatively fixes are outlined below.

Glacier thickness and bed topography: One of the major parameters required for defining the SHAKTI model domain is bed elevation. Because there have been no ice-penetrating radar systems and no borehole measurements at Shishper, we must rely on bed topographies that are inferred from surface elevation (DEMs). Here, we have used the data from [49], which uses a mathematical inversion to infer bed thickness for all glaciers contained in the Randolph Glacier Inventory. By using [99]'s database which uses an improved mass balance technique to estimate glacier thickness, we may be able to obtain more accurate bed topography data for the model domain. This can give us more accurate estimates of ice overburden pressure and effective pressure, which affect channel formation and uplift. Another way to improve bed topography estimates is to use DEMs from previous decades during which currently glaciated areas were uncovered; however, as the advancing of the glacier tends to alter the topography of the land beneath it, this method should also be undertaken with careful thought.

Updates to glacier outlines: The glacier outline we used to define the domain is from RGI 6.0, which is based on glacier outlines from 2016 [39]. It was well established by [26] that Shishper's terminus advanced significantly during the 2017-2019 surge. Using an updated glacier outline taken from more recent remote sensing imagery will allow greater accuracy of the model.

Calibration of gap height: As mentioned earlier, we made a mistake during the winter spin-up (5.3.1) by not letting it run until the gap height reached equilibrium. As a result, we saw that the average gap height across the mesh gradually went upwards every year, where we expected to see the gap height return to the same winter conditions every year. In the next iteration of this work, we will allow the gap height to equilibrate fully by running the winter spin-up for 300 to 400 days. This will allow us to more accurately model the seasonal evolution of the subglacial drainage system.

Mesh resolution sensitivity study: All model variables in SHAKTI are discretized by mesh element. In [147], a mesh resolution sensitivity analysis was carried out to understand how the mesh resolution affects channel formation. They found that as the mesh element size increased from 50m to 200m, fewer channels formed and the degree of branching decreased. In any finite element simulation, it is ideal to obtain results that are independent of mesh size. We will conduct a similar mesh resolution sensitivity study by varying the mesh resolution (50 meters is used in these experiments) between 20 and 100 m and examining how model outputs change.

6.2.2 The Effects of Liquid Precipitation

In this work, the only hydrological input we considered was surface (ice) melt and basal melt. Since we had prescribed the surface velocity to be 0, the basal melt was quite small. However, along with the ERA5 ice ablation data, we have also been provided with daily estimates of liquid precipitation. By adding liquid precipitation to the hydrological inputs, we can consider the impact of the Asian monsoon season on the subglacial system. In particular, we can examine the role of large storms in triggering channel formation or bed lubrication at various times during the intersection of the melt season and the monsoon season.

In the next simulation, we will consider SHAKTI's "englacial input" term to be comprised of ice melt plus liquid precipitation. Besides shedding light on the role of rainfall, the increased volume of meltwater will be more faithful to the physical system. Thus, we will be able to get a fuller picture of surge and GLOF events that occurred at the site. We can also evaluate whether rain can be a trigger for surges or GLOFs by doing a similar comparison as 6-13.

6.2.3 Proximal Lake Drainage

In our Discussion, we have dedicated a great deal of attention to analyzing conditions that may be conducive to surges. However, the more urgent threat originating from Shishper is GLOF. The 2017-2019 surge caused Shishper to dam the meltwater flow from the neighboring glacier, Muchuwar, creating an ice-dammed lake (Figure 6-14). This ice-dammed lake has drained through Shishper's terminus multiple times, resulting in evacuations of the village of Hassanabad and Hunza Valley [20].

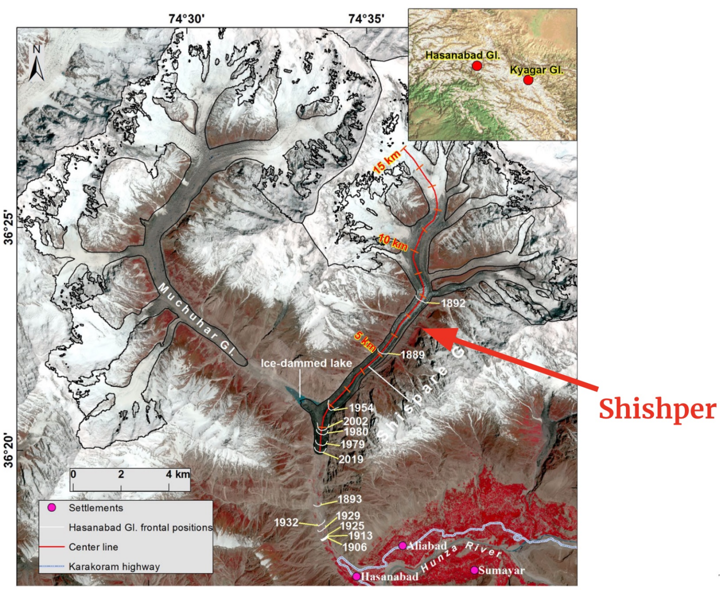

Figure 6-14: *Historical position of Shishper Glacier [26]*

These lake drainages range from less dramatic flows in the winter to full-scale GLOFs in the spring which cause significant damage to homes, crops, and infrastructure. According to [20], farmers and residents of Hassanabad are constantly living in fear of their homes and fields being suddenly swept away by a lake drainage. To gain a better understanding of how GLOFs may affect the drainage network at Shishper's terminus, we can simulate a buildup of hydrostatic pressure at the left side of Shishper. This is done by assigning a Dirichlet (fixed) water pressure boundary condition along the bottom left stretch of the domain outline where the proximal lake comes in contact with the glacier.

6.2.4 Moulin (Supraglacial Lake) Drainage

The SHAKTI model allows the user to specify where and when supraglacial lakes (moulins) empty into the subglacial environment. As mentioned earlier, [90] found that surface uplift following the drainage of a supraglacial lake (moulin) can allow us to infer the transmissivity of the subglacial system. This could provide a potential avenue for model validation in the future. Multispectral satellite imagery is commonly used to identify glacial lakes and surface melt by calculating an index called NDWI (Normalized Difference Water Index) [154]. We can use this method to track supraglacial melt on Shishper (analysis of satellite imagery may also help us to parametrize water storage in heavily crevassed/damaged areas). In tandem, we can track uplift over the glacier using DEMs to see where and when "water blisters" form. We can then model any observed lake drainages and uplift in SHAKTI and use the relaxation of the blisters to infer the subglacial hydraulic transmissivity using methods from [90]. This can be compared to transmissivity K modeled in SHAKTI. The results of a study like this would provide a degree of much-needed model validation for SHAKTI and would help to put constraints on various modeling parameters.

6.2.5 Coupled Hydrology and Ice Dynamics Simulations

As established earlier, we have neglected many feedbacks and components of the subglacial system by not coupling the hydrology model to the full ice-dynamics model. There are many complicated feedbacks between surface velocity, basal velocity, frictional heating, cavity opening, and till shearing that can affect both surge behavior and the behavior of the subglacial system as well.

To shed light on surge dynamics, we will set up a fully (two-way) coupled hydrology and ice dynamics model. We would like to model the 2017-2019 surge with greater complexity, showing how the terminus of the glacier moving approximately 1500 m (1.5 km) may have affected the subglacial hydrology system and vice versa. Since the initial SHAKTI model development [146], two-way coupling has been developed and tested for various example domains. This is an exciting new development and will lead to many extensions of this work.

6.2.6 Contributions to Early Warning Systems

Early warning systems (EWS) allow residents downstream GLOF-prone lakes extra time to evacuate when glacial lakes are going to flood dramatically. Although there are currently many EWSs already installed and functional in the Himalayas, GLOFs are still regarded as extremely difficult to predict, and none of the existing EWSs can automatically trigger an alarm ahead of time. Our ability to make accurate predictions of these hazards is currently quite limited, as we have not been able to fully understand why lake drainages and GLOFs occurred in the past. However, the preliminary simulation outputs from this work are very promising. If future work allows us to reliably simulate historic lake drainages and GLOFs, running this model in real-time may give us the ability to create a more accurate early-warning system to alert local people of high-danger days when a glacial lake is at risk of outburst.

6.3 Conclusion

The SHAKTI model is one of the first subglacial hydrological models that has the ability to model a continuum of distributed to channelized drainage systems. It is the only subglacial hydrology model with these capabilities that is connected to ISSM and can be readily coupled with sophisticated ice dynamics simulations. This work is the first application of SHAKTI to a glacier in High Mountain Asia, an area of critical importance for the future of humanity (Chapter 1). The Karakoram region is notorious for its glacial surges and GLOFs, and there are large populations living just downstream of these hazards. Since there have been limited observational studies done in this area, models like this, which are informed by remote sensing and local knowledge, can provide a crucial step forward in understanding these glaciers.

In this study, we find that the SHAKTI model produces results that are feasible and intuitive according to previous observational and modeling studies of subglacial hydrology in Greenland, Antarctica, and other alpine environments. We find that the subglacial drainage system transitions from a closed, distributed system in the winter to an efficient, channelized one by late summer. We also show that a single channel remains year-round at the bottom trunk of the glacier and provides the basis for the rest of the drainage system during the melt season. There have also been many studies that suggest various mechanisms for the starting and stopping of surge behavior. Subglacial hydrology is likely a major factor behind surge behavior, but our model results suggest that subglacial hydrology is just one of multiple important mechanisms that start and stop surges. It is also more likely that subglacial hydrology plays more of a role in terminating surges than starting them.

This initial study on Shishper Glacier provides the nucleus for a new branch of research in modeling alpine glaciology. This will allow us to gain a better understanding of how Himalayan glaciers respond to climate change, and will help illuminate why still poorly-understood phenomena such as GLOFs and glacial surges occur. In the Karakoram specifically, it has been commonly observed for valley glaciers to dam other valley glaciers, resulting in proximal glacial lakes such as the one seen at Shishper and Muchuwar [66]. Our work here can then provide a useful framework for studies on many similar glacier systems in the Karakoram range, amplifying the impact of this work.

Chapter 7

Supplementary Materials

This folder contains 10 videos, each containing 6 months of simulated basal flux (shown in log scale). The suffix "_1" denotes the first 183 days of the year (January-June). The suffix "_2" denotes the second 182 days of the year (July-December). 2020_2_basalflux.mp4 contains 183 days because 2020 was a leap year. The sampling rate for each of the videos is 1 per day, meaning that each frame of the video represents 1 day of simulated time.

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